CHAPTER 4

The Atmospheric Circulation System

Key Questions

- Why does air move?
- Are the movements of the winds random across Earth's surface, or do they follow regular patterns?
- What implications do these circulatory systems have for global climate?
- What other factors govern the geographic and seasonal distributions of temperature and rainfall?

Chapter Overview

Earth's climate is a central theme of ours. We focus on the role climate plays in the Earth system and explain how Earth's climate works, how climate has changed through time, and how it may change in the future. An important element of Earth's climate is the atmospheric circulation. In Chapter 3 we described the global energy budget and showed that if we average the radiation fluxes around the globe and over a few years there is a balance. Earth emits as much energy as it receives, aside from the issue of anthropogenic increases in the greenhouse effect. If we look at regions smaller than the globe and over time periods of less than a year, however, the situation is very different. There is a significant imbalance in the distribution of energy at various latitudes: The tropics receive a surplus of radiative energy, whereas the poles run a deficit. This imbalance causes an equator-to-pole temperature gradient that results in density and pressure differences in the atmosphere. The density and pressure differences cause air to move in a global-scale pattern of wind belts, which are modified by Earth's rotation and by the distribution of land and water. The net effect is to restore the latitudinal energy balance by moving surplus energy away from the tropics to cancel out the deficit at the poles. In the process, energy is used to evaporate water from the land and ocean surfaces, water vapor is carried by wind, and energy is released when the vapor condenses to form clouds. Thus, there are close interactions between the transport of energy and of water by means of circulating air. In other words, Earth's atmospheric circulation has a direct impact on the global distributions of temperature and precipitation.

THE GLOBAL CIRCULATORY SUBSYSTEMS

Anyone who has felt wind blow, watched clouds move, and seen rain fall is aware that large parts of the Earth system are in constant motion. Even the continents and oceans, despite their apparent permanence, are continuously moving. The island of Iceland in the North Atlantic, for example, is spreading, and its two sides are moving away from each other fast enough to be measured by today's instruments. Although these movements may sometimes appear random, they form part of a well-ordered circulation of energy and matter throughout the Earth system.

Like the circulatory system of humans (part of the cardiovascular system), Earth's circulatory subsystems work to maintain the planet in a thermal and chemical balance. The human circulatory system transports dissolved gases, nutrients, and hormones throughout the body; carries away waste products; helps regulate the acidity of body fluids; and is a vital part of the body's thermoregulatory system, carrying warm blood from one area to another. Although the human circulatory system is not an exact analogy to Earth's circulatory subsystems, these systems do have much in common. Essential gases and nutrients are transported throughout the Earth system, and waste products are removed from their area of production. All of Earth's circulatory subsystems act in some way to help regulate the global temperature: The winds and ocean currents redistribute the energy received from the Sun, and the motions of the solid Earth redistribute carbon and help regulate the CO_2 level of the atmosphere. The circulations within the solid Earth are discussed in Chapter 7; here and in Chapter 5 we examine those circulations that occur within the fluid part of the Earth system: the atmosphere and oceans.

The purpose of this chapter is to describe the major characteristics of the atmospheric circulation, to explain why they occur, and to illustrate the way in which they affect the transport of energy and materials around the globe. In Chapter 3 we described the energy input and output from the Earth system as a whole; now we take that system apart and examine some of its internal workings—specifically those related to climate. In doing so, we have two primary objectives. The first is to explain why weather and climate vary across the globe. The second is to emphasize that because of the internal workings of Earth's climate system, the response to global-scale processes and changes may not be uniform around the globe. Organized movements of the atmosphere occur over many different time and space scales. These movements range from centimeter-scale swirls, or *eddies,* to global-scale motions of the wind belts. All of these are important in one way or another, but we limit our discussion to processes that are global in extent and that have the greatest influence on the transport of energy and mass through the Earth system. One of Earth's most important constituents is water. Cycling continuously among the atmosphere, the oceans, and the land surface, water carries with it energy, dissolved nutrients, and other matter—all vital for maintaining an environment suitable for life.

In the same way that the functioning of the cardiovascular system in the human body ultimately depends on the ability of the heart to keep pumping, the functioning of Earth's circulatory subsystems rely on several different pumps. Each of these pumps drives a different circulatory mechanism, and each works at a different speed. Over shorter time scales (years to decades), the most important pump is found in the tropical oceans. This pump is responsible for the movements of the air and the surface ocean over most of the globe. The energy source that drives this pump is radiation from the Sun. Over longer time scales (about 1,000 years), a second pump drives the deep-ocean circulation (Chapter 5). The ultimate energy source is again the Sun. The pump operating over the longest time scales (millions of years) is radioactive decay and the production of heat in Earth's interior. This pump causes the movements of the continents, which we discuss in Chapter 7.

All these circulation subsystems play a vital role in the operation of the Earth system. We know that in humans the cardiovascular system can maintain stability only if the blood keeps moving. Similarly, the Earth system can maintain stability only as long as its circulatory subsystems continue to function. The long-term pump (that is, the processes of internal heat production and plate tectonics) ceased to function on Mars. As we will see in a later chapter, the planet's inability to support an environment suitable for life may be at least partly a result of the failure of this circulation mechanism.

THE ATMOSPHERIC CIRCULATION

Recall from Chapter 3 that the troposphere is the lowermost layer of the atmosphere. Most of the processes we are interested in take place in the troposphere, so we limit our discussion here to that layer. Although the circulation of the stratosphere plays a role in the depletion of stratospheric ozone, we will save that discussion for Chapter 17.

The Movement of Air

Air moves over Earth's surface because there are horizontal differences in pressure. Air also moves vertically either because it is forced to rise mechanically (e.g., when it encounters a mountain range) or because there are changes in *buoyancy.* **Buoyancy** is the tendency of an object to float in a fluid. Buoyancy is controlled by differences in *density* between the object and the fluid, where density is given by the mass of a substance within a unit volume. (The greater the mass within a given volume, the greater the density.) Ultimately, all of these horizontal and vertical movements (except those due to mechanical forces) can be attributed to differences in temperatures across the globe. To explain how these movements occur, we need to understand how pressure and density are related to temperature.

VERTICAL MOVEMENT It is easiest to picture these relationships by thinking of vertical and horizontal movements separately. Imagine the situation with a hot-air balloon. Remember from Chapter 3 that heating causes molecules to move faster. In this case, the faster the air molecules move, the more they collide with each other and with the interior of the balloon. These collisions exert a force (i.e., air pressure) on the interior surface. If the balloon was a fixed container (one that could not expand), there would be an increase in the air pressure within the container. Thus we see a connection between the temperature and pressure of a gas: As the temperature increases, the pressure increases. But the balloon *is* expandable. As the pressure starts to increase, the air pushes outward on the interior of the balloon, causing it to expand. So in a balloon, it is the volume rather than the pressure that increases (see the Box "A Closer Look: The Relationships between Temperature, Pressure, and Volumes—The Ideal Gas Law").

If we begin with the balloon partially inflated, it will contain a certain number of air molecules. As these are heated, the number of molecules does not change, but they move faster, increasing the pressure on the interior of the balloon; this causes the balloon to expand. We now have the same number of air molecules as before (the mass [m] hasn't changed), but they occupy a greater volume (V). This means that the density of the air (ρ) must decrease ($\rho = m / V$). Because the air in the balloon is less dense than the air surrounding it, the balloon becomes *positively buoyant,* and it rises. The balloon will continue to rise until the density of the air outside the balloon matches that inside (*neutral buoyancy*). If the air in the balloon is more dense than the surrounding air, the balloon would have *negative buoyancy,* and it would sink. Exactly the same processes occur in the atmosphere when we heat the surface below a parcel (column) of air. The surface heats the parcel of air at the bottom of the column; the air parcel expands, its density decreases, and the parcel rises through the air column. Cooling a parcel of air higher in the column causes it to become more dense than the surrounding air, and the parcel sinks.

HORIZONTAL MOVEMENT How do horizontal movements occur? We saw that warmer air has a lower density than cooler air. If we consider two adjacent columns of air, one warmer than the other, the cooler column would have a greater density and higher pressure than the warmer column. This difference in pressure would cause the air to move horizontally from the region of higher-pressure cool air to the region of lower-pressure warmer air—the air moves down the pressure gradient. The atmospheric pressure is a force (**F**) determined by the mass (m) of the air column and the acceleration (**a**) due to gravity (remember Newton's second law of motion: $\mathbf{F} = m\mathbf{a}$). Averaged globally and through time, the atmosphere exerts a pressure of 1013 mbar on Earth's surface. Thus 1013 mbar is considered to be one atmosphere of pressure (1 atm). The pressure decreases as you rise up in the atmosphere (because there is less air above you) until, at the top of the atmosphere, the pressure reduces to zero. The actual pressure recorded at any point on the surface or in the atmosphere, however, can be highly variable under different conditions of elevation and temperature. So, for adjacent columns of air with similar volumes, the colder high-density air has a higher atmospheric pressure. Hence, as you see on weather charts, air flows from high pressure regions to regions of low pressure.

A CLOSER LOOK

The Relationships between Temperature, Pressure, and Volumes—The Ideal Gas Law

We can see from the text discussion that we can define specific relationships among the temperature (T), pressure (P), and volume (V) of a gas. (In this case, volume refers to the space occupied by a fixed mass of gas molecules; density is the inverse of volume for a fixed mass, so we need not consider it separately here.) First, we know that if the temperature is held constant, then an increase in gas pressure results in a decrease in volume, and a decrease in pressure results in an increase in volume. In other words, pressure is inversely proportional to volume: The product of pressure and volume is a constant. Mathematically, we can write

$$P_{initial} V_{initial} = P_{final} V_{final}, \text{ (Boyle's law)}$$

where $P_{initial}V_{initial}$ is the product of the initial pressure and volume, and $P_{final}V_{final}$ is the product of the final (new) pressure and volume. This relationship, known as **Boyle's law,** was discovered in 1662 by the British chemist Robert Boyle.

If, instead, gas pressure is held constant, then we know that an increase in temperature results in an increase in volume and that a decrease in temperature results in a decrease in volume. In other words, volume is directly proportional to temperature: The quotient of volume and temperature is a constant. Mathematically, we can write

$$\frac{V_{initial}}{T_{initial}} = \frac{V_{final}}{T_{final}}, \text{ (Charles's law)}$$

where $V_{initial}/T_{initial}$ is the quotient of the initial volume and temperature, V_{final}/T_{final} is the quotient of the final volume and temperature, and temperature is in kelvins. This relationship, known as **Charles's law,** was discovered in 1787 by the French physicist Jacques Charles.

All gases are found to behave the same way over a wide range of conditions, and Boyle's law and Charles's law can be combined to give the ideal gas equation:

$$PV = mRT$$

where m is the mass and R is a constant called the gas constant for 1 kg of gas (the value of R depends on the particular gas concerned).

It is important to note that these relationships apply to idealized gases, that is, gases in which there are no attractive forces between molecules. In reality, gases may not respond to environmental changes exactly as described here. Nevertheless, these relationships do represent a very close approximation of how gases behave.

From this discussion we can establish two important points that will help explain why air moves:

1. Air tends to move from an area of higher pressure to an area of lower pressure until the two pressures are equalized. In other words, air (wind) will move horizontally in the lower troposphere from higher to lower pressure. Pressure differences among air masses are typically related to the distribution of surface temperatures.
2. If an air mass is heated until its density is lower than that of its surroundings, the lower-density air will rise. This phenomenon is a form of convection. (We discussed this phenomenon in a more general sense in Chapter 3 when we demonstrated the convection of a fluid that is heated from below.) Conversely, if an air mass is cooled until its density is higher than that of the underlying air, it will sink. This phenomenon is referred to as **subsidence.**

The Driving Force: The Global Energy Distribution

We learned in Chapter 3 that the average global temperature is determined by the balance between the solar energy absorbed by Earth and the infrared radiation emitted to space. However, neither the radiation received from the Sun (our primary energy source) nor the infrared emission from Earth is distributed uniformly across Earth's surface. The incoming solar energy varies with latitude and with season, whereas the outgoing terrestrial radiation depends on the temperature of the surface and atmosphere at each location.

The distribution of the incoming solar radiation changes with latitude as a result of the change in surface area presented to the Sun's rays as Earth's surface curves (Figure 4-1). The energy from the Sun radiates outward in all directions; however, by the time the Sun's rays reach Earth, they are essentially parallel to each other. This means that the flux of solar energy passing perpendicularly through the plane *A-B* in Figure 4-1 will be the same at any point. For example, the three "beams" in the diagram are equal in solar flux when they pass through the plane. Because of the curvature of Earth, however, when these beams reach the top of Earth's atmosphere, the same amount of light is spread over a much larger area at the poles than at the equator. Consequently, each square meter of surface receives proportionately less energy at the higher latitudes, and the incoming solar flux thus decreases from the equator toward the poles. (Recall the lightbulb and sheet of paper experiment in Chapter 3.)

The solar radiation absorbed at the surface follows the same general pattern, although the actual amount absorbed varies with cloud cover and atmospheric absorption. This equator-to-pole gradient in the energy absorbed at the surface exerts a primary control on Earth's climate. Figure 4-2 shows this gradient (solid curve) as a function of latitude (i.e., the amount averaged around each latitude band). As we might expect from the previous discussion, the maximum absorbed solar energy is found in the tropics, and the available solar energy decreases rapidly as we move toward the poles. This gradient in absorbed solar energy is the single most important control on temperature. More energy is generally available at the equator than at the poles, so we can assume that temperatures should be highest in the tropics and lowest at high latitudes. Figure 4-2 also shows the latitudinal distribution of infrared radiation emitted from Earth to space (dashed curve). The higher emissions in the tropics are a result of the high surface temperatures there and the correspondingly high temperature in the middle troposphere,

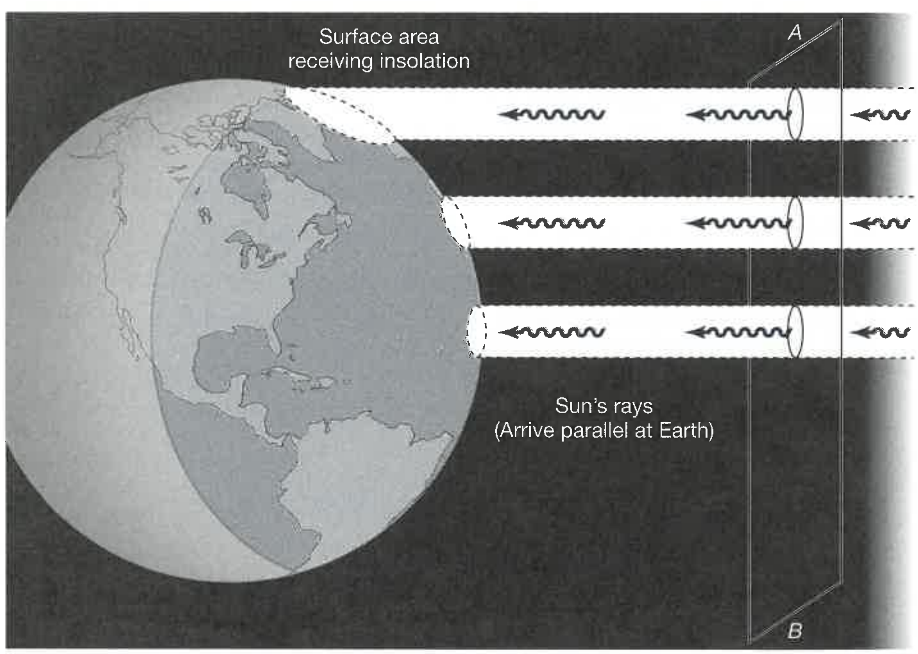


FIGURE 4-1 Variation of incoming solar energy with latitude. The radiation reaching Earth is spread over larger and larger areas as we move from the equator to the poles. Each square meter of the surface receives proportionately less energy as we move to higher latitudes. (*Source:* From R.W. Christopherson, *Geosystems: An Introduction to Physical Geography,* 3/e, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

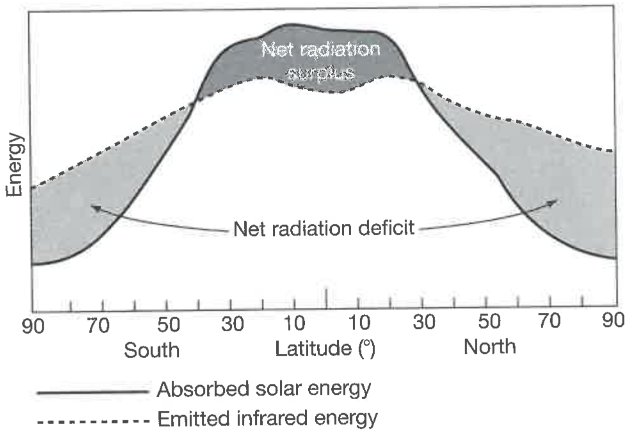


FIGURE 4-2 The distribution of absorbed solar and emitted infrared radiation with latitude. There is a surplus of energy in the tropics, where incoming radiation is greater than outgoing, and a deficit at high latitudes, where more radiation is emitted than is received.

from whence the outgoing radiation is emitted. Again, you can refer back to the discussion of the IR flux–temperature feedback described in Chapter 3.

The difference between the incoming solar radiation and the outgoing terrestrial radiation is referred to as *net radiation*. Referring again to Figure 4-2, note that the energy absorbed exceeds the energy emitted in the tropics (net radiation is positive); near the poles, the reverse is true (net radiation is negative). This distribution of available energy is a permanent feature of Earth's climate system. The gradient seems to imply that the tropics should get warmer while the poles get progressively colder. Clearly this does not happen; other processes must be operating to ensure an energy balance at each latitude. In reality, the latitudinal energy gradient produces atmospheric temperature and density differences that force the atmosphere to circulate, carrying warmer air toward the poles and colder air toward the equator. These circulations move energy from regions where there is a surplus to regions where there is a deficit. Most of what we experience as weather and climate is this response of the atmosphere to the unequal latitudinal distributions of energy.

The General Circulation of the Atmosphere

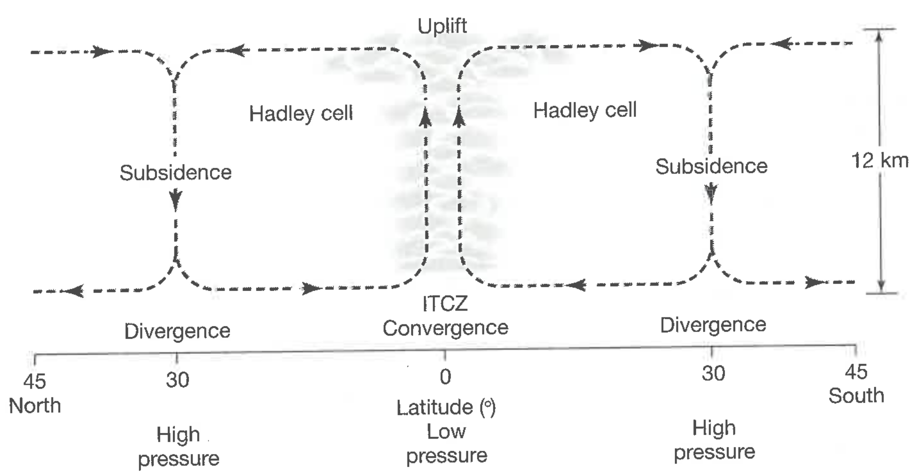
From our description of the energy distribution and our discussion of how air movements occur, we can build a picture of what we would expect the global-scale circulation of the atmosphere to look like. This circulation involves several characteristic features that we will discuss in turn. Taken together, these circulation features represent a negative feedback loop as the atmosphere responds to the temperature gradient by transferring energy latitudinally to reduce the gradient and restore an energy balance. The continuous addition of energy from the sun of course means that the energy distribution is never balanced.

CONVERGENCE We begin with the heating in the tropics. The large solar input to the tropics heats the surface (primarily ocean), which in turn heats the overlying air. As we saw earlier, when heated from below, air will rise by convection. The tropical air near the surface rises, creating a low-pressure region there. But we saw that air tends to move horizontally from an area of higher pressure to an area of lower pressure. Thus, the rising air is replaced by surface air moving equatorward into the region of low pressure from regions of higher pressure (Figure 4-3). The merging of air masses that are moving inward toward a low-pressure region is called **convergence.** The converging air masses that meet at the tropics and rise make up the **intertropical convergence zone (ITCZ).**

The surface heating produces evaporation in addition to convection. As the convecting air rises, it cools, and the evaporated water (water vapor) in the convecting column condenses to form clouds. As a consequence, the ITCZ is characterized by extensive areas of cloud cover and heavy precipitation. We talk more about evaporation, condensation, and rainfall later in the chapter.

DIVERGENCE The top of the troposphere, located at about 12–15 km in the tropics, forms a barrier to further uplift.

FIGURE 4-3 Convergence, divergence, and the Hadley circulation in the tropics. There is a Hadley cell on either side of the intertropical convergence zone (ITCZ), located over the equator. Rising air in the ITCZ is replaced by inflowing air (convergence) at the surface. Outflowing air (divergence) in the upper troposphere sinks at about 30° N and 30° S, completing the circulations in the two cells.



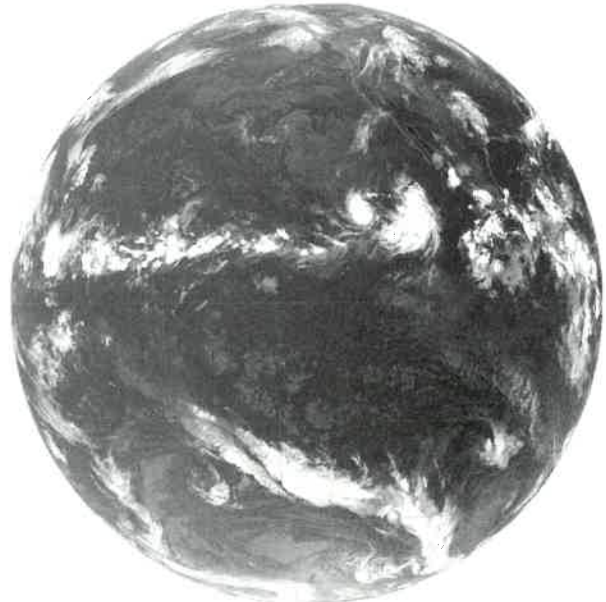


FIGURE 4-4 Satellite image of the eastern Pacific and Central America. These images are obtained from geostationary satellites, which orbit over the equator at an altitude of about 35,000 km and at an orbital speed that keeps pace with Earth's rotation; thus the satellite appears to remain stationary over the same spot on the equator. This image was obtained by the National Oceanographic and Atmospheric Administration (NOAA) during Northern Hemisphere summer. A line of convective clouds marks the ITCZ just north of the equator. The clear areas to the north and south of the ITCZ mark the descending arms of the Hadley cells. (*Source:* NOAA.)

Remember that temperatures generally increase in the stratosphere and that the higher temperatures produce a stable structure that limits convection from below. The air that rises in the ITCZ, upon reaching this barrier, is forced to diverge poleward. **Divergence,** in this case, refers to the movement of air outward from a region in the atmosphere. This poleward-moving air subsides at about 30° N and 30° S latitude, replacing the air that is moving equatorward at the surface (Figure 4-3). The air warms as it sinks, which prevents condensation from occurring and clouds from forming. As a result, these regions are characterized by clear skies and low rainfall amounts. If you check an atlas, you will find that such areas coincide with some of the world's largest deserts (e.g., the Sahara and Arabian deserts and the Great Australian Desert). The subsiding air also leads to an area of high pressure and divergence at the surface.

FIGURE 4-5 Convective towers in the ITCZ. Solar heating evaporates large amounts of water from the tropical oceans. The air cools and condenses as it rises, releasing the energy used for evaporation as latent heat. The release of latent heat in these convective towers is the pump that drives the Hadley circulation. (*Source:* NASA.)

HADLEY CIRCULATION This pattern of air movement, with convergence occurring in the tropics and divergence and subsidence some 30° away in one large convection cell, is called **Hadley circulation.** This circulation pattern was named for George Hadley, the British meteorologist who first explained the phenomenon. The convection cells on either side of the equator, referred to as *Hadley cells*, represent the dominant north–south mode of circulation between 30° N and 30° S latitude. Note, however, that the Hadley cells—and the ITCZ—are not continuous around the globe. The circulation takes place in individual cells of rising and subsiding air, and the pattern is further broken up by land–ocean contrasts. The ITCZ is most obvious in the Atlantic and Pacific oceans and is readily observed in satellite images such as that shown in Figure 4-4. The large-scale circulation in Southeast Asia and the Indian Ocean is dominated by the monsoon, which is described later in this chapter.

The convection cells in the ITCZ result directly from surface heating in the tropical oceans. In fact, although solar heating provides the fuel for tropospheric circulation, the actual pump that drives the circulation is the release of latent heat during convection. (We discussed latent-heat release in Chapter 3.) The energy of solar radiation, used to evaporate water from the ocean surface, is converted to latent heat, and the latent heat is released to the atmosphere in huge towers of convective cloud clusters within the ITCZ (Figure 4-5). It is this release of latent heat that pumps the air around each Hadley cell.

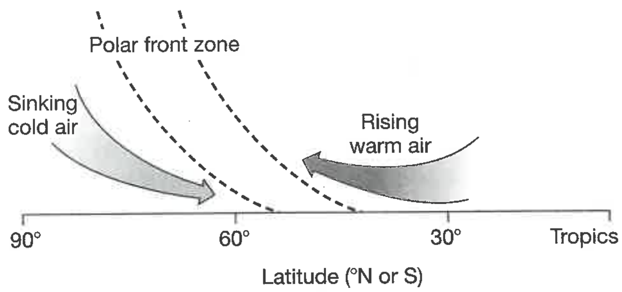


FIGURE 4-6 Mixing of air in the midlatitudes. The lower-density warm air from the tropics rises above the colder air moving equatorward from high latitudes. These contrasting air masses do not mix very easily. This zone is characterized by large temperature contrasts over very short distances.

MIDLATITUDE AND HIGH-LATITUDE CIRCULATION

Thus far we have discussed the atmospheric circulation only between the equator and 30° N or S. What about atmospheric circulation from there to the poles? The very low temperatures at the poles, particularly in winter, result in increased air density near the surface and, thus, in higher pressures than occur in the tropics. The higher density and pressure lead to divergence and a general movement of cold air outward at the surface, that is, toward the equator. The divergence is accompanied by subsidence from above. The equatorward-moving cold air meets the warm air moving poleward from the subtropics, producing a zone of steep temperature gradients called the **polar front zone** at approximately 60° N and S latitude. The two air masses do not mix easily: The warm air is less dense than the cold air, which therefore sinks below the warm air when the two air masses meet (Figure 4-6). The polar front zone, therefore, slopes poleward with increasing altitude in the atmosphere. Note that, because of dynamic processes that come into play when air moves over a curved surface, this frontal zone forms a wavelike structure around the hemisphere. The actual latitude at which the front is located, therefore, varies from place to place.

When we put Figures 4-3 and 4-6 together, we see an alternating pattern of northward- and southward-moving air at the surface (Figure 4-7). Such north–south movement is called *meridional* circulation. If we look at Figure 4-7 from above, we might expect to see a general pattern of surface winds such as those depicted in Figure 4-8. We would expect surface winds to blow out of the high-pressure zones at the poles and at about 30° N and S, and to blow toward the low-pressure zones at the equator and at about 60° N and S. The actual pattern, however, is more complicated because winds tend to blow in east–west directions as well. Indeed, the east–west motions are considerably greater than the north–south motions. We know that differences in solar heating cause the equator-to-pole movement we have been discussing. What causes the east–west movements?

The Coriolis Effect

East–west movements of surface winds are the result of the Coriolis effect. The **Coriolis effect** (named for Gaspard Gustav de Coriolis, the French mathematician who in 1835 proposed that the concept applies to surface winds) is the apparent tendency for a fluid (air or water) moving across Earth's surface to be deflected from its straight-line path. (Some texts refer to a *Coriolis force* in relation to this effect. This force, however, is only an apparent force due to the observer's frame of reference, not a real force due to an identifiable source, such as the gravitational pull of a planet.) Viewed from Earth, a north–south moving object appears to be deflected to the east or west. Viewed from space, the same object is in fact seen to move in a straight line. The apparent curve that we see is the result of our frame of reference—we normally view the object's movement from *within* the system.

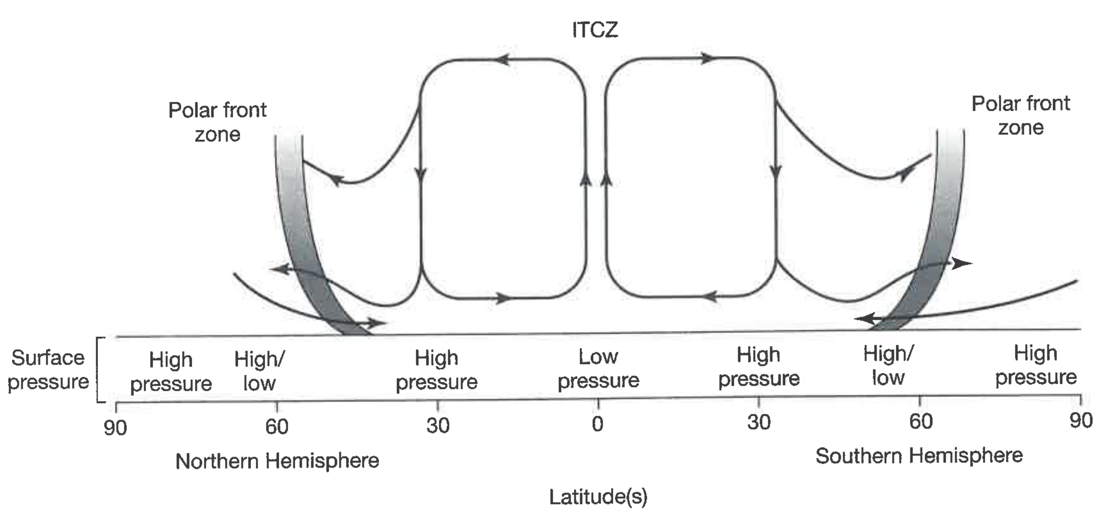


FIGURE 4-7 The north–south (meridional) circulation of the troposphere. The tropical circulation is dominated by the Hadley circulation, whereas midlatitude circulation and weather are controlled by the location of the polar front zone and the mixing of cold polar air with warm air from the tropics.

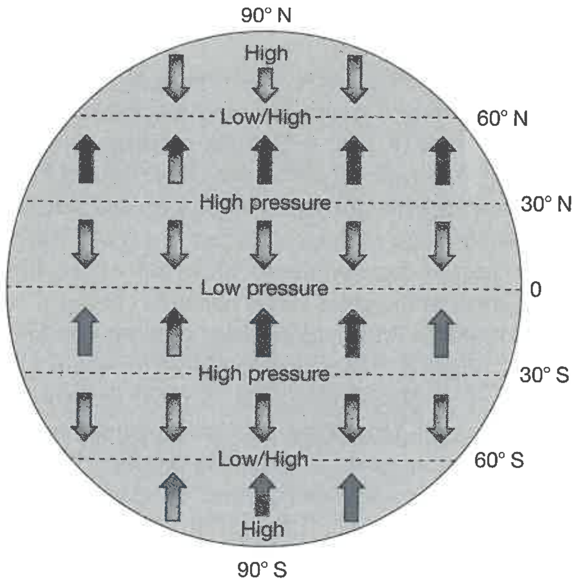


FIGURE 4-8 A possible model of the surface winds obtained by plotting, on a globe, the pattern of surface winds that would be deduced from Figure 4-7. Surface winds blow out of the high-pressure zones at the poles and at 30° N and 30° S and blow toward the low-pressure zones at the equator and in the midlatitudes.

The Coriolis effect applies to any object moving on a rotating body. To visualize this, let us first consider Earth rotating on its axis. The two longitudinal lines in Figure 4-9a represent the distance moved in a given time interval, and the arrows represent the rotation speed of Earth's surface at different latitudes over that interval. The speed of rotation is greatest at the equator (approximately 464 m/sec), and it decreases as we move north—or south—until it becomes zero at the poles.

Now imagine an object, such as an air mass, that is apparently stationary at a point on Earth's surface. Although this air mass is not moving relative to the surface, it is traveling eastward at Earth's rotation rate for that location. (For example, an object that is stationary at any one of the points marked by the left-hand edges of the rotation arrows in Figure 4-9a would, to an observer in space, appear to move along the arrow's path.) Because the rotation rate changes with latitude, an air mass moving northward from the equator—from point *A* to point *B* in Figure 4-9b—will appear to curve off to the right of its straight-line path, arriving at *X* rather than at *B*′. Why does this happen?

In the time it would take the air mass to travel from *A* to *B*, Earth rotates from *A* to *A*′ (and *B* moves to *B*′). Remember that the air mass at *A* is moving not only northward but also eastward at Earth's speed of rotation (represented by the distance *A*-*A*′). As long as it is between the equator and point *B*, the air mass is moving from west to east faster than Earth is rotating at *B*. Thus, the air mass will "gain" on the ground below it and will arrive at point *X* instead of at *B*′. Although it is difficult to visualize, the air does in fact move in a straight line; if we were watching from space rather than from Earth, that is what we would see.

We have seen how an air mass (or any object) moving northward in the Northern Hemisphere is deflected to its right. Following the same reasoning, we can see that an air mass moving southward in the Northern Hemisphere also curves to its right (relative to the direction of initial movement), because now the air mass is moving eastward more slowly than Earth's surface immediately underneath. The easiest way to keep track of this is to think of the deflection direction relative to the direction of initial motion of the object—it is always to the right of the direction of initial motion in the Northern Hemisphere and to the left in the Southern Hemisphere.

What happens when the initial direction of movement is due east or due west? As it happens, the Coriolis effect still comes into play, but for a different reason. When an object is set in motion along a circular path, *centrifugal force*—another apparent force—tends to push the object away from the center of rotation. (This is the same phenomenon that forces your car off the road if you try to turn a corner too fast.) If an air mass in the Northern Hemisphere is moving eastward faster than Earth is rotating at that latitude, that air mass will experience an apparent centrifugal force that pushes it directly away from Earth's spin axis.

We can break down this apparent force into two components: one component that is acting perpendicular to the

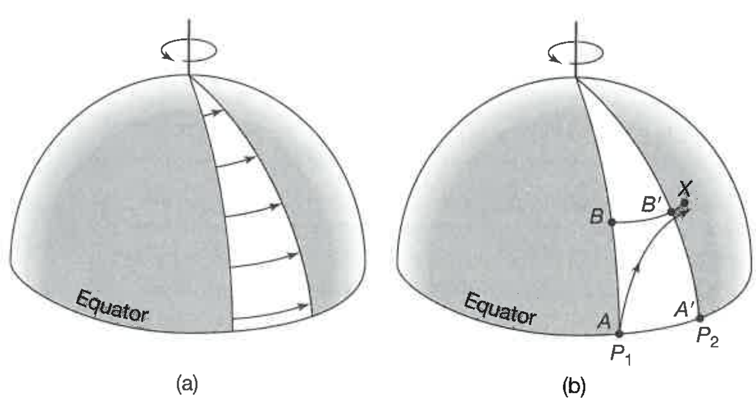


FIGURE 4-9 The Coriolis effect. (a) As Earth rotates, the speed of the surface is greatest at the equator and is zero at the poles. (b) At the equator, Earth has rotated from *A* to *A*′. Points *A* and *B* have moved to *A*′ and *B*′. Air initially moving from *A* toward *B* would actually curve to the right and arrive at point X.

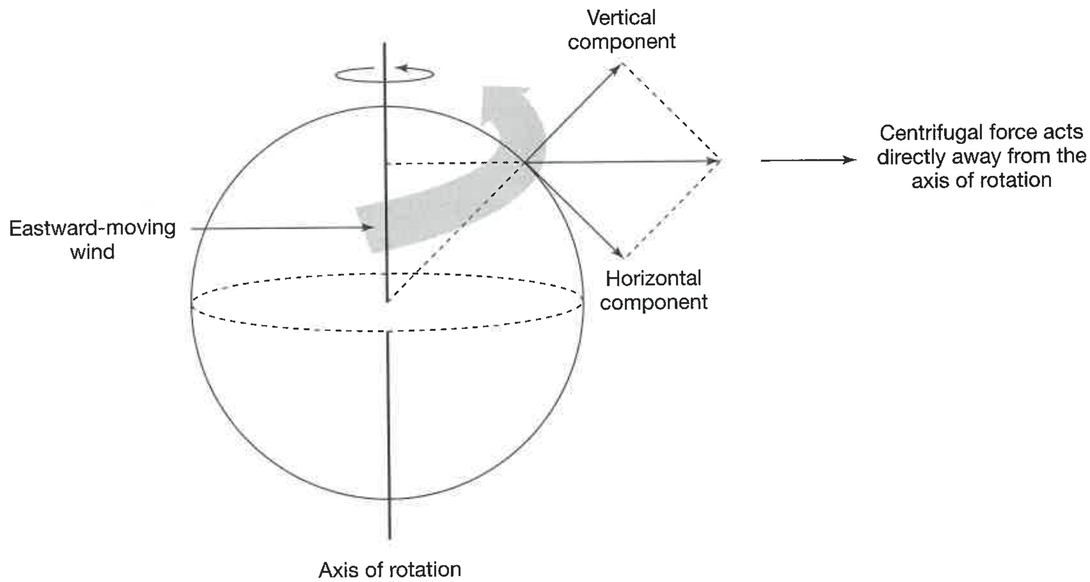


FIGURE 4-10 The Coriolis effect produced by the centrifugal force acting on eastward- or westward-moving winds.

surface, and one that is horizontal (parallel) to the surface (Figure 4-10). For an eastward-moving wind in the Northern Hemisphere, the horizontal component is to the south; the wind would curve to the south, or to the right. For a westward-moving wind, the horizontal component is to the north; the wind still curves to the right. In the Southern Hemisphere, the effects would be opposite: Both eastward-moving and westward-moving winds would curve to the left.

Although these descriptions of the deflection effect for north–south and east–west moving objects appear to be very different, mathematically the deflections are identical, and both are referred to simply as the Coriolis effect. The Coriolis effect increases as the speed of the object increases. And whereas the speed at which Earth's surface is moving due to its rotation is zero at the poles and a maximum at the equator, the Coriolis effect is zero at the equator and increases with latitude. The only place on Earth's surface where the Coriolis effect does not come into play is at the equator. An air mass moving eastward or westward around the equator is not deflected from its original path. Such an air mass is not changing latitude, so there is no Coriolis effect due to the difference in rotation rate with latitude. Nor is there a horizontal component to the centrifugal force, so again there is no Coriolis effect.

Distribution of Surface Winds

We can now modify the simplistic pattern of northward-moving and southward-moving surface winds in Figure 4-8 to obtain the more realistic pattern of surface winds shown in Figure 4-11. There the winds are deflected to the right and left of the paths of initial motion in the Northern and Southern hemispheres, respectively. This deflection of the winds due to the Coriolis effect gives rise to easterly winds at high latitudes. Meteorologists refer to winds in terms of the direction from which they blow. In other words, an "easterly" wind is a wind that blows from east to west. The midlatitudes are characterized by westerly flow, and the tropics by easterly winds called the *northeast* and *southeast trade winds.* Winds at the equator tend to be highly variable in direction. This region where winds are light and frequently change direction is referred to as the *doldrums*.

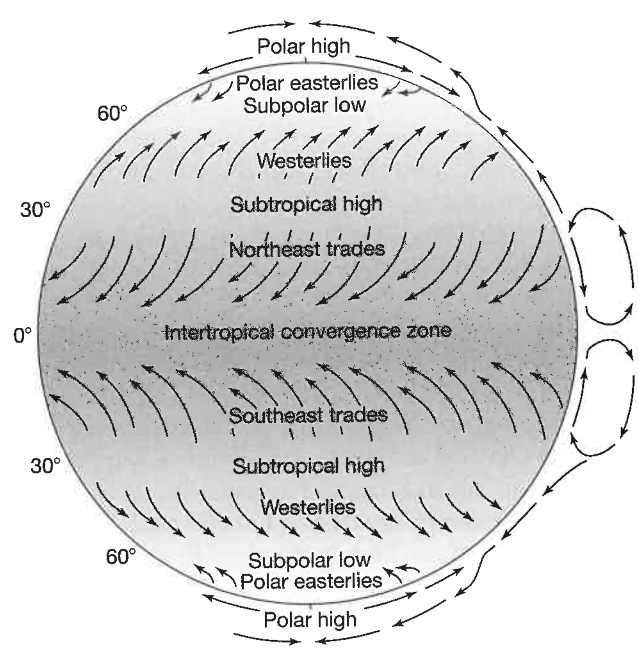


FIGURE 4-11 The pattern of surface winds. This shows the same general pattern of winds as Figure 4-8, but the wind directions have been changed to include the deflection due to the Coriolis effect. (*Source:* From T. McKnight, *Physical Geography: A Landscape Appreciation*, 6/e, 1999. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

Considerations of pressure differences, buoyancy, and the Coriolis effect have led us to a good first approximation of the general circulation of the troposphere. This pattern, however, is still a little too simplistic. In reality the indicated winds, for example, do not blow continuously, and they are not continuous around the globe. As we noted earlier, uplift in the ITCZ takes place in clusters of convective cells rather than in two giant cells, one on each side of the equator. The rising air moves poleward and, under the influence of the Coriolis effect, turns to the right in the Northern Hemisphere (and to the left in the Southern Hemisphere). It thus becomes a westerly flow in the upper troposphere. Some of this air subsides near 30° N or S latitude to form the subtropical high-pressure belt (Figure 4-11), but the subsidence too is concentrated in localized areas. The locations of these high-pressure systems vary with season, although they are always found in these approximate locations. The trade winds blow from the equatorward side of these semipermanent high-pressure systems. Similarly, poleward-moving air from the subtropical high-pressure zone curves due to the Coriolis effect, producing a generally westerly flow in the midlatitudes. The actual flow pattern, however, is highly variable from day to day.

The pressure and wind patterns in the midlatitudes depend on the location of the subtropical highs as well as on the distribution and movement of temporary areas of high or low pressure that form in association with the steep temperature gradients in the polar front zone. Small areas of low pressure (on the order of 1000 km wide) form in this zone in part due to the surface-temperature gradient, but also because of dynamic processes occurring higher in the troposphere. As air blows into these regions, it curves to the right (in the Northern Hemisphere), producing a localized circular flow pattern referred to as *cyclonic flow.* Air flowing out of a high-pressure region (referred to as an *anticyclone*) will also curve to the right in the Northern Hemisphere, creating an *anticyclonic,* or clockwise, flow. (The direction of air flow around cyclones and anticyclones is reversed in the Southern Hemisphere.)

Low-pressure systems that form outside the tropics are referred to as *extratropical cyclones.* The circular flow mixes warm air from the equatorward side of the system with cooler air from the high latitudes. As we noted earlier, the warm and cool air masses do not mix easily; hence these systems are characterized by extensive uplift as the warmer, less dense air rises above the cooler and denser air mass. As we will see later in the chapter, this rising air results in the formation of rain or snow. These circulation features move along the polar front, bringing low pressure as they move over a region, which is then replaced by higher pressure as they move past. These transient high-pressure and low-pressure systems are characteristic features of midlatitude climates and account for much of the day-to-day variability in weather in these regions. At high latitudes, the polar easterlies are most clearly developed in winter (when the coldest surface temperatures occur and when this area of low temperatures is most extensive). However, low-pressure systems from the midlatitudes often migrate into the polar regions in summer, breaking down the surface high-pressure systems and disrupting the easterly flow.

UPPER-LEVEL FLOW Referring back to Figures 4-7 and 4-8, we see that the pressure distribution at the surface results in alternating regions of poleward and equatorward moving air. At higher levels, however, Figure 4-7 suggests that all of the air is moving from the ITCZ toward the poles. Why the difference?

Think back to the description of the latitudinal energy balance and the discussion of pressure, temperature, and volume relationships. On the large scale, we have a planet where the troposphere has warm air in the tropics and relatively cooler air at the poles. As the warmer air expands and the cooler air contracts, then the depth of the troposphere changes with latitude (which is also suggested by Figure 4-7). Consequently, we have the situation shown in Figure 4-12a. As the troposphere is thicker in the tropics than at the poles, then the change in pressure with height must be slower in the tropics, as shown in Figure 4-12b. If we join these tropical and polar pressure surfaces, we get the situation shown in Figure 4-12c. Notice we have given a slight slope to the pressure surfaces to reflect the decreasing temperatures toward higher latitudes.

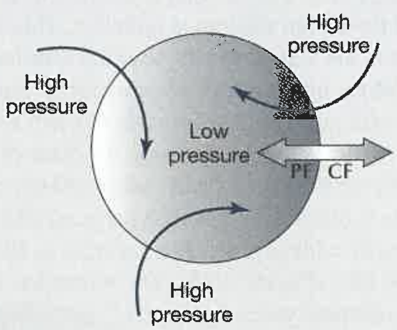
Now, away from the surface where local temperature and pressure changes can be large, if we take any line of constant altitude in the diagram (e.g., A-A' or B-B') we see that at any point along these lines the pressure is always higher on the equatorward side than the poleward side. Air will flow down the pressure gradient from high to low pressure and, on average, the flow at higher levels in the troposphere is from the tropics to the pole. You can imagine that the wind speed will be greatest where the pressure gradient is the steepest, which happens in the upper troposphere in the midlatitudes. Belts of high wind speeds that we see in this location are referred to as *jet streams* (indicated by the J in Figure 4-12c).

What if we take a horizontal view? We know that if the air is moving poleward, and thus changing latitude, it must come under the influence of the Coriolis effect. Consequently, the air will curve to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. In other words, there will be a westerly component to the flow in both cases. The force that is pushing the air down the pressure gradient is referred to as the pressure gradient force. This force is balanced by the Coriolis effect—such that the air actually flows at right angles to the gradient and the resulting movement is referred to as the *geostrophic wind* (Figure 4-13). Other forces (centripetal and centrifugal forces) come into play if the air is following a curved trajectory around centers of higher or lower pressure, where the greater the curvature, the more the flow departs from being geostrophic. Friction also plays a role, with its greatest effect being close to the surface. Friction between

A CLOSER LOOK

How Hurricanes (Tropical Cyclones) Work

Tropical cyclones (also called *hurricanes* when they occur in the Atlantic and *typhoons* when they occur in the western Pacific oceans) are strong low-pressure centers accompanied by powerful, and sometimes devastating, winds. These storms begin their lives as relatively mild, low-pressure centers overlying tropical oceans. Air from the surrounding higher-pressure regions moves inward toward the center of low pressure—at which point several different forces come into play. The flow around a hurricane is determined by the pressure gradient force and centripetal acceleration, balanced by the Coriolis effect, causing the flow to be to the right of the pressure gradient in the Northern Hemisphere and to the left in the Southern (see Box Figure 4-1). Note that because tornadoes are too small for the Coriolis effect to come into play, tornadoes can, in fact, rotate in either direction. The balance of forces causes the air to spiral inward in a counterclockwise direction (Northern Hemisphere) or a clockwise direction (Southern Hemisphere). The swirling winds increase the rate of evaporation of seawater, adding large amounts of moisture to the air. Because the air is converging near the surface within the hurricane, it rises in the clouds circling the storm (except, curiously, in the very center—the eye—where the air is descending). The rising air cools, causing this moisture to condense as rain and, at the same time, liberating its latent heat. This strengthens the storm, speeding up the winds and



BOX FIGURE 4-1 Schematic diagram of a Northern Hemisphere hurricane. The circle represents an isobar (a line of constant pressure). The arrow labeled PF represents the pressure force; the arrow labeled CF represents the Coriolis force. The winds blow counterclockwise turning to the right as they move from high to low pressure.

drawing in more air from outside the storm; this air rises in the clouds and the process continues in a positive feedback loop. As long as the system remains above warm tropical waters, it acts as a self-sustaining heat engine. Eventually, though, it will either move poleward, where the surface water is colder, or it will encounter land. In either case, the warm ocean water that was providing it with energy is no longer present, and the hurricane dissipates. The storm can also weaken over warm water if another weather system overpowers it.

Several conditions are necessary in order for tropical cyclones to form. One, already noted, is that they require warm ocean temperatures—generally in excess of 26 to 27°C. Because the Coriolis effect is necessary to cause the air to start to turn, then storms can't form within about 5° latitude of the equator (remember the Coriolis effect is zero at the equator and increases toward the poles). There needs to be low vertical wind shear. Wind shear occurs when the horizontal winds higher in the troposphere are moving with a much different speed or direction than the winds nearer the surface. When this happens, it tends to tear the storm apart. It also helps to have a rapid decrease in temperature with height, producing positive buoyancy and increasing convection, and also to have relatively high water vapor content in the lower and middle levels of the atmosphere. This keeps the air near the saturation vapor pressure as it rises and maximizes the amount of latent heat release (see the discussion of vapor pressure later in this chapter). Tropical cyclones also need some initial atmospheric disturbance to get them started. This can arise in various ways, such as convergence in the intertropical convergence zone, old frontal boundaries that have moved down over the subtropical oceans and, in the case of Atlantic hurricanes, the initial storm development can be caused by small disturbances in the easterly flow off Africa (referred to as easterly waves). Conditions are best for hurricanes in late summer and fall when the ITCZ has moved furthest from the equator and when sea-surface temperatures are at their warmest. Apart from the Coriolis effect, however, all these other characteristics can vary over time (from year to year or over longer time periods) causing variations in the number and intensity of hurricanes that occur.

the moving air and the surface reduces the wind speed and counters some of the Coriolis effect. In this case, rather than flowing at right angles to the gradient (like the geostrophic wind), the air flows down the gradient at an angle less than 90°. This is why surface winds spiral into low-pressure centers and out of high-pressure centers rather than circling them (which would be the case if the flow was geostrophic). This spiraling effect is clearly seen in Figure 4-20.

Moving back to the upper troposphere, where friction is not an important issue, we see that the flow is close to being geostrophic. It flows normal to the gradient, but it follows a wavelike path around the globe (Figure 4-14). For dynamical reasons that are more complex than we want to get into here, when a fluid moves over a rotating surface, it follows a wavelike trajectory—in this case curving toward the equator and back toward the poles in several large waves that extend around each hemisphere. These waves were first described mathematically by Carl G. Rossby in 1938, and are now referred to as *Rossby waves*. The number of waves, their location, and how well developed they are varies from day to day. For those of us living

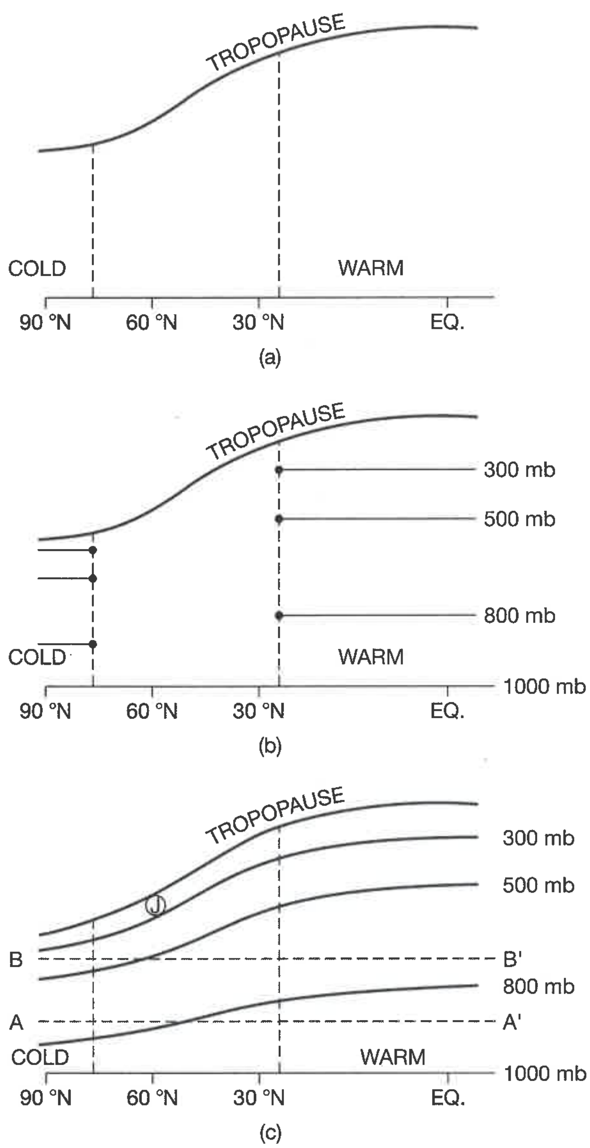


FIGURE 4-12 Pressure change with height in the troposphere. The higher temperatures in the tropics causes the air to expand, raising the height of the tropopause compared to the poles (a). As the atmospheric pressure is the same in both regions, the decrease in pressure with height must be slower in the tropics compared to the poles (b). If the tropical and polar pressure surfaces are joined, we see the steepest pressure gradients in the midlatitudes. At any height in the atmosphere, the pressure is higher in the tropics than it is at the poles (c). As wind speeds are greatest where the pressure gradient is steepest, the highest wind speeds (the jet streams) will be found high in the troposphere in the midlatitudes—the *J* in Figure 4-12c.

in the midlatitudes, these waves steer the low- and high-pressure systems that produce our day-to-day weather. This is why television weather forecasts often show you the locations of the jet streams—as they control the path that low-pressure systems will follow.

SEASONAL VARIABILITY The simple pattern of global winds shown in Figure 4-11 is also modified by seasonal variations. As well as changing with latitude, the distribution

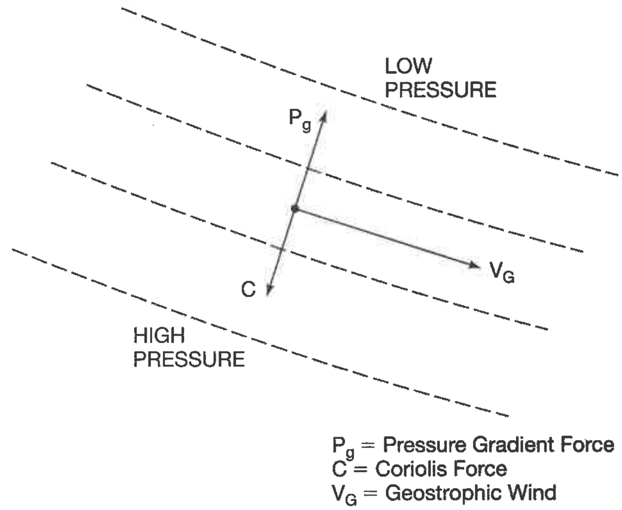


FIGURE 4-13 The geostrophic wind results from the balance of the pressure gradient force and the force due to the Coriolis effect. The dashed lines are isobars—lines of equal pressure. The pressure gradient force (P_g) acts perpendicular to the isobars in the direction of the low pressure. This is balanced by the force due to the Coriolis effect (C), resulting in the geostrophic wind (V_G) blowing parallel to the isobars and to the right of the pressure gradient force (in the Northern Hemisphere).

of solar energy varies with the seasons. Figure 4-15 shows the seasonal pattern of Earth's orbit around the Sun. The time when Earth is closest to the Sun is referred to as *perihelion;* Earth is farthest from the Sun at *aphelion.* This difference in distance from the Sun affects the seasonal distribution of temperature. More important to seasonality is Earth's *tilt,* or **obliquity.** Obliquity refers to the angle of Earth's spin axis relative to a line drawn perpendicular to the plane of the planet's orbit around the Sun. Each planet has a different angle of tilt. Earth's axis is tilted 23.5° from the perpendicular. On human time scales, the obliquity remains constant as Earth revolves around the Sun (Figure 4-15). On somewhat longer time scales, the obliquity varies by about ±1° (see Chapter 14).

For six months of each year, the Northern Hemisphere faces the Sun and the Southern Hemisphere faces away; for the other six months, it is the Southern Hemisphere that faces the Sun while the Northern Hemisphere faces away. The hemisphere that faces the Sun receives much more solar energy than does the other hemisphere. It is this factor that determines the seasons. Consider the June 21 solstice (the day with the longest period of sunlight in the Northern Hemisphere) in Figure 4-15. Imagine Earth spinning around its axis. The North Pole will remain in sunlight the whole time, while the South Pole will remain in darkness. The opposite is true at the solstice on December 21 (the day of shortest sunlight duration in the Northern Hemisphere). The result is six months of sunshine and six months of darkness at the poles.

The greatest heating occurs where the Sun is directly overhead. Due to Earth's obliquity, the latitude at which this occurs varies continuously throughout the year, from 23.5° S (called the Tropic of Capricorn) on December 21 to 23.5° N (called the Tropic of Cancer) on June 21. There

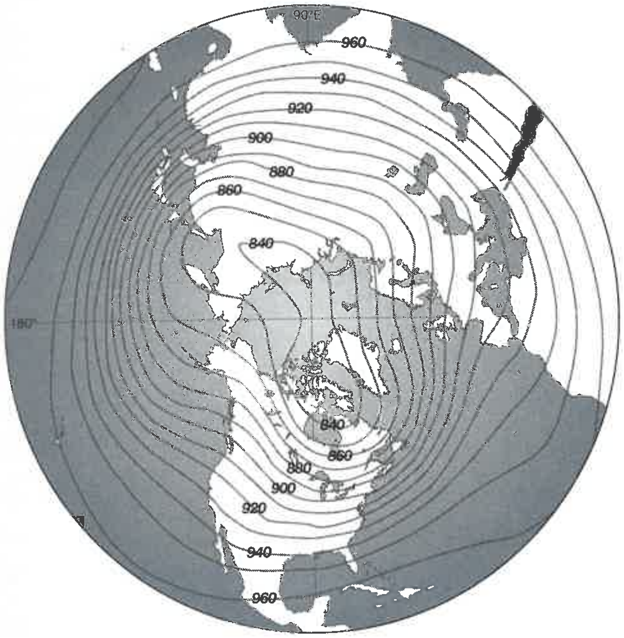


FIGURE 4-14 Northern Hemisphere mean January 300-mbar geopotential heights. The heights are in decameters (1 decameter = 10 m). Essentially, the map shows the height of the 300-mbar surface. The surface slopes down from the tropics to the Arctic and follows a wavelike structure around the hemisphere. (*Source:* Map provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, from its website, http:/ /www.cdc.noaa.gov/)

are also two days on which the Sun is directly overhead at the equator—the *vernal equinox,* March 20, and the *autumnal equinox,* September 22. On these dates, daytime and nighttime are equal in length. (These dates vary by a day or so because the calendar is adjusted every four years to reconcile it with Earth's revolution around the Sun.) Notice, though, that within a narrow band near the equator the Sun is always close to being overhead. Hence there is a large input of solar radiation to the tropics at all times. The difference between daytime and nighttime temperatures in this region, in fact, is usually much greater than the seasonal difference there.

The seasonal variability in incoming energy shifts the atmospheric circulation patterns northward and southward as the seasons change (Figure 4-16). The hemisphere experiencing summer has less of a temperature gradient between the tropics and the pole than does the opposite hemisphere. The fact that the Sun shines continuously for six months at each pole compensates for the fact that the poles do not receive as much solar energy per unit area as do the tropics. This reduced temperature gradient weakens the strength of the atmospheric circulation. At the same time, because the Sun is directly overhead somewhere away from the equator, the maximum solar energy is directed somewhere poleward. Consequently, the steepest temperature gradients are shifted toward the poles, and the circulation patterns are also shifted poleward. In the winter

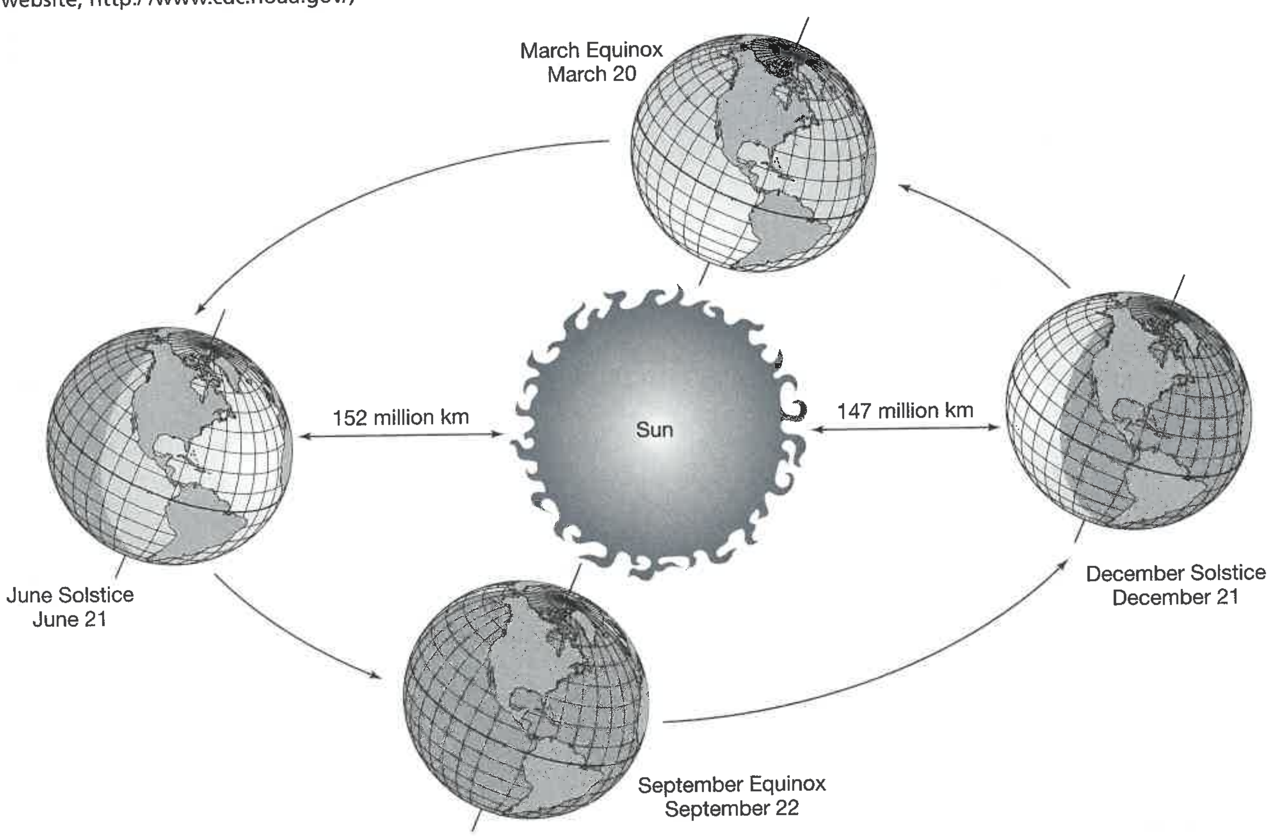


FIGURE 4-15 The seasons. The seasons are controlled by Earth's obliquity and Earth's orbit around the Sun. The hemisphere that is "tilted" toward the Sun experiences summer while it is winter for the hemisphere that is "tilted" away from the Sun. The equinoxes (when the Sun is directly overhead at the equator) mark the transition seasons: fall and spring. (*Source:* From T. McKnight, *Physical Geography: A Landscape Appreciation,* 6/e, 1999. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

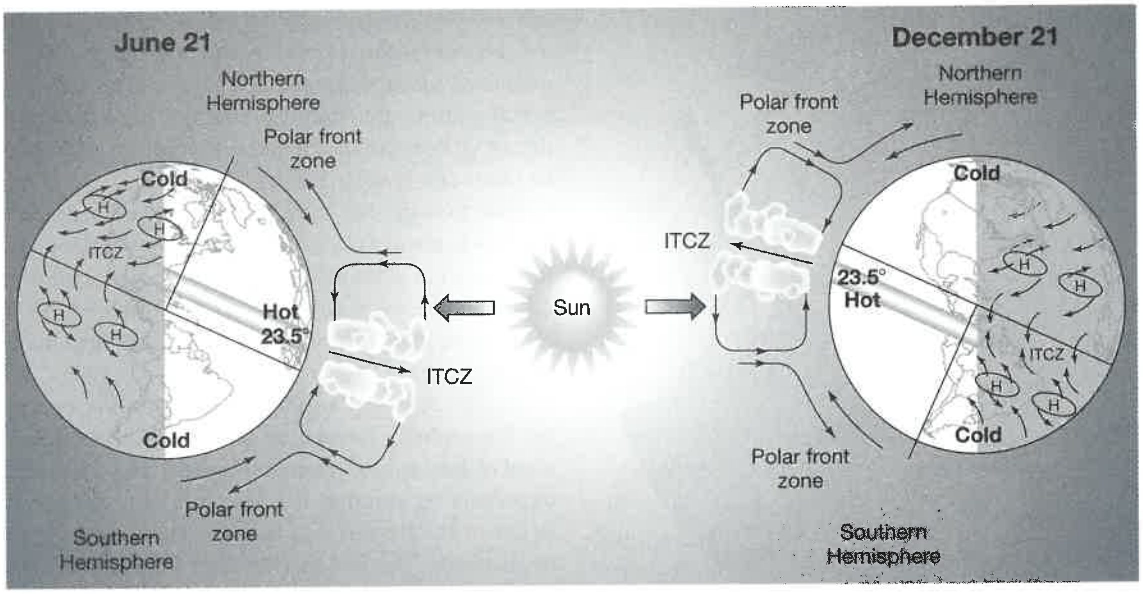


FIGURE 4-16 Seasonal migration of the atmospheric circulation patterns. The ITCZ is found in the summer hemisphere, where the circulation is weaker and the patterns are shifted toward the pole. The subtropical high-pressure cells that mark the descending arms of the Hadley circulations are denoted by H.

hemisphere (with six months of darkness at the pole), the equator-to-pole temperature gradient is much stronger and the steepest gradients are shifted equatorward. As a result, the atmospheric circulation is more intense and the circulation patterns are shifted toward the equator.

The ITCZ also moves northward and southward as a result of these seasonal shifts in insolation. The ITCZ will reach its maximum northward location late in the Northern Hemisphere summer. (There is a time lag in all of these shifts between the time that the solar heating occurs and the resulting shift in the circulation pattern.) It will then migrate southward, crossing the equator in the fall and reaching its most southern location late in winter (Southern Hemisphere summer). The upper tropospheric circulation is similarly affected, with more intense wind speeds in winter and the jet streams shifting north and south with the seasons. Note that when the ITCZ is located over the equator, the poleward-moving air in the upper troposphere produces westerly winds in both hemispheres. As the ITCZ shifts northward, however, the southward-moving air will turn to the right, producing easterly winds in the equatorial region, before curving back to the west when it reaches the Southern Hemisphere. The same thing will happen in reverse when the ITCZ is south of the equator. So, while most of the upper tropospheric flow is westerly, there are frequently narrow bands of easterly winds in the equatorial regions.

GLOBAL DISTRIBUTIONS OF TEMPERATURE AND RAINFALL

In the first part of this chapter, we described some of the main features of the global-scale atmospheric circulation. For the remainder of the chapter, we look at the effect this circulation has on other parts of the Earth system, specifically, the global temperature and rainfall distributions.

As we have learned, the ultimate cause of the atmospheric circulation is the distribution of available energy. More interesting for our purposes is the fact that the interaction between temperature and circulation is not a one-way process. As we indicated earlier, the circulation itself is an important component of Earth's thermoregulatory system, transporting energy (heat) from areas where there is a surplus to areas where there is a deficit.

The transport of water is also strongly affected by the atmospheric circulation. The distribution of water about the globe is important in that organisms require a sufficient supply of water to maintain life. That distribution is also important for the transport of dissolved materials. As we will soon see, evaporation and precipitation are strongly influenced by temperature and, therefore, by the distribution of energy. Furthermore, the transport of water in its various forms (liquid, water vapor, and ice) also modifies the temperature distribution by affecting the radiation budget (Chapter 3) and thus feeds back to affect the circulation. Hence, we see that temperature, precipitation, and the atmospheric circulation are all closely linked and that interactions and feedbacks exist among all three of these components of Earth's climate.

Land–Ocean Contrasts

Beyond the latitudinal distribution of energy, global temperature patterns are also strongly influenced by the distribution of land and ocean. Recall from Chapter 2 (Table 2-1) that the albedo of the ocean surface is considerably lower than the albedo of most land surfaces. In consequence, oceans absorb more of the available solar energy than do land surfaces at the same latitude.

Land and ocean surfaces also behave very differently in what they do with that energy. An ocean surface rapidly transfers heat downward by turbulent mixing and to the atmosphere above by convection. Part of the contrast between land and ocean is due to differences in their thermal properties. The land surface rapidly loses heat to the atmosphere by convection, but it transfers heat downward relatively slowly by conduction. How easily this transfer occurs depends on the physical and chemical properties of the material; the rate at which this occurs is described by its *thermal conductivity*. More formally, thermal conductivity is the rate at which heat energy passes through a column of material that has a temperature gradient along the column of 1 K, or 1°C, per meter. Water has a high thermal conductivity, whereas land surfaces have low thermal conductivities. Furthermore, we can consider the *heat capacity* of the two types of surfaces. Heat capacity at constant volume is the energy required to raise the temperature of a unit mass of a substance by 1 K or 1°C without changing its volume. In other words, heat capacity is a measure of how much energy must be added to an object to change its temperature. The heat capacity of water is about three to four times that of dry soil. Thus, the input of a given amount of energy will raise land temperatures much more than it will raise sea-surface temperatures.

More important is that the ocean surface transfers heat rapidly downward by turbulent mixing. As we will see in Chapter 5, the surface layers of the ocean are well mixed; when the ocean surface is heated, the heat is mixed downward within the surface layers. The amount of material that must warm (or cool) is very large, so the temperature change is very slow. A further factor is differential absorption of the two surfaces. Whereas all the solar radiation falling on the land surface is reflected or absorbed right at the surface, some of the solar radiation falling on the ocean penetrates and is absorbed below the surface. Hence, energy is transferred downward even more rapidly in water than on land.

THE SEA BREEZE Putting all this together, we see that with equal amounts of incoming energy, land surfaces will heat up much more rapidly than do ocean surfaces but will also cool down much more rapidly once the input of energy is reduced. Land surfaces heat up quickly during the day and cool quickly at night, whereas ocean surfaces warm slowly in the day, and the temperature drops very little at night. The sea breeze that occurs near coastlines is a direct result of this diurnal variability. The heating of the land surface during the day warms the overlying air and gives rise to small areas of low pressure and uplift; cooler temperatures over the ocean result in relatively higher pressures and subsidence of the cooled air above (Figure 4-17). Air flows down the pressure gradient (from the area of higher pressure to the area of lower pressure), creating onshore flow from ocean to land. At night this temperature structure breaks down, and the atmospheric circulation weakens. If the land cools sufficiently, the circulation pattern may reverse.

CONTINENTALITY We can see the same effect on a larger scale in terms of seasonal climate variability. As we noted earlier, the seasonal variation over midlatitudes and high latitudes is much greater than in the tropics. This variability is also much greater over land surfaces than over the oceans because of their different thermal characteristics. This property is referred to as *continentality*. The more continental the climate, the more it is characterized by seasonal temperature extremes. Land surfaces are much warmer than ocean surfaces in summer and much colder in winter. The effect of continentality on global temperatures is visible in Figure 4-18c. The greatest seasonal variability

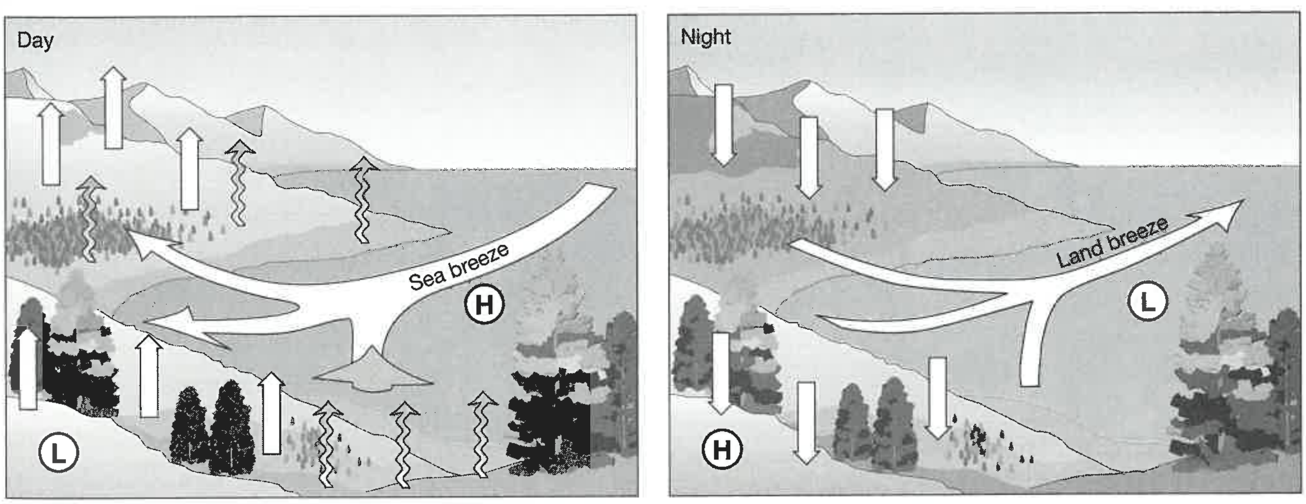


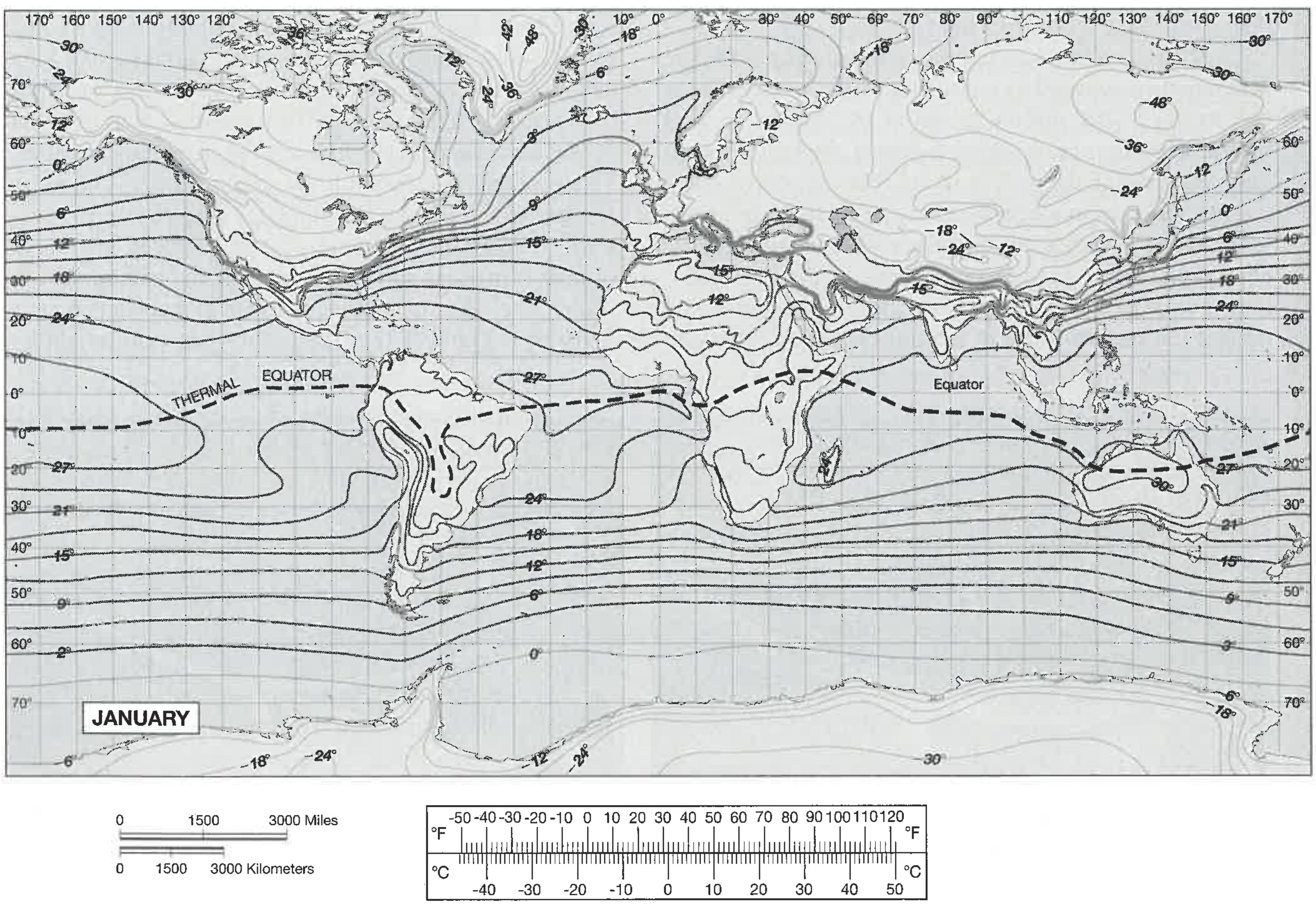
FIGURE 4-17 The sea breeze. The heating of the land during the day causes localized convection with low pressure at the surface. This convection establishes a pressure gradient from the ocean toward the land that results in onshore wind. At night, the rapid cooling of the land (relative to the ocean) causes this circulation to break down and may even reverse the flow, causing the wind to blow from the land toward the water. (*Source:* From T. McKnight, *Physical Geography: A Landscape Appreciation*, 6/e, 1999. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

is found in the interior of large continental masses, and the lowest variability is over the tropical oceans. The oceans provide a moderating effect in coastal regions that reduces the temperature extremes. The temperature difference between ocean and land surfaces also affects the mean sea-level pressure distribution (again feeding back to affect the circulation). Figure 4-19 shows the average atmospheric pressure that would be found over land if the land surfaces were at sea level. In Northern Hemisphere winter (Figure 4-19a), the very cold surface temperatures of interior North America and Asia cool the lower layers of the atmosphere, producing high surface pressures over those landmasses. The North Atlantic and North Pacific are characterized by low-pressure zones produced by the low-pressure systems forming along the steep temperature gradient in the polar front zone. In summer this gradient decreases (Figure 4-19b). The low-pressure zones are less well developed and are displaced poleward, the subtropical highs expand, and the continental regions of high pressure are replaced by regions of low pressure.

The Southern Hemisphere (Figures 4-19a and 4-19b), with its huge expanse of ocean and very little land area, presents a less complicated picture. The midlatitude air flow is much more *zonal*—that is, the air circulation more closely follows lines of latitude—than it is in the Northern Hemisphere. The seasonal temperature change causes the air flow patterns to shift north and south slightly and causes pressure gradients to change, but the distribution of land and oceans produces much less variability of winds around the Southern Hemisphere than around the Northern Hemisphere.

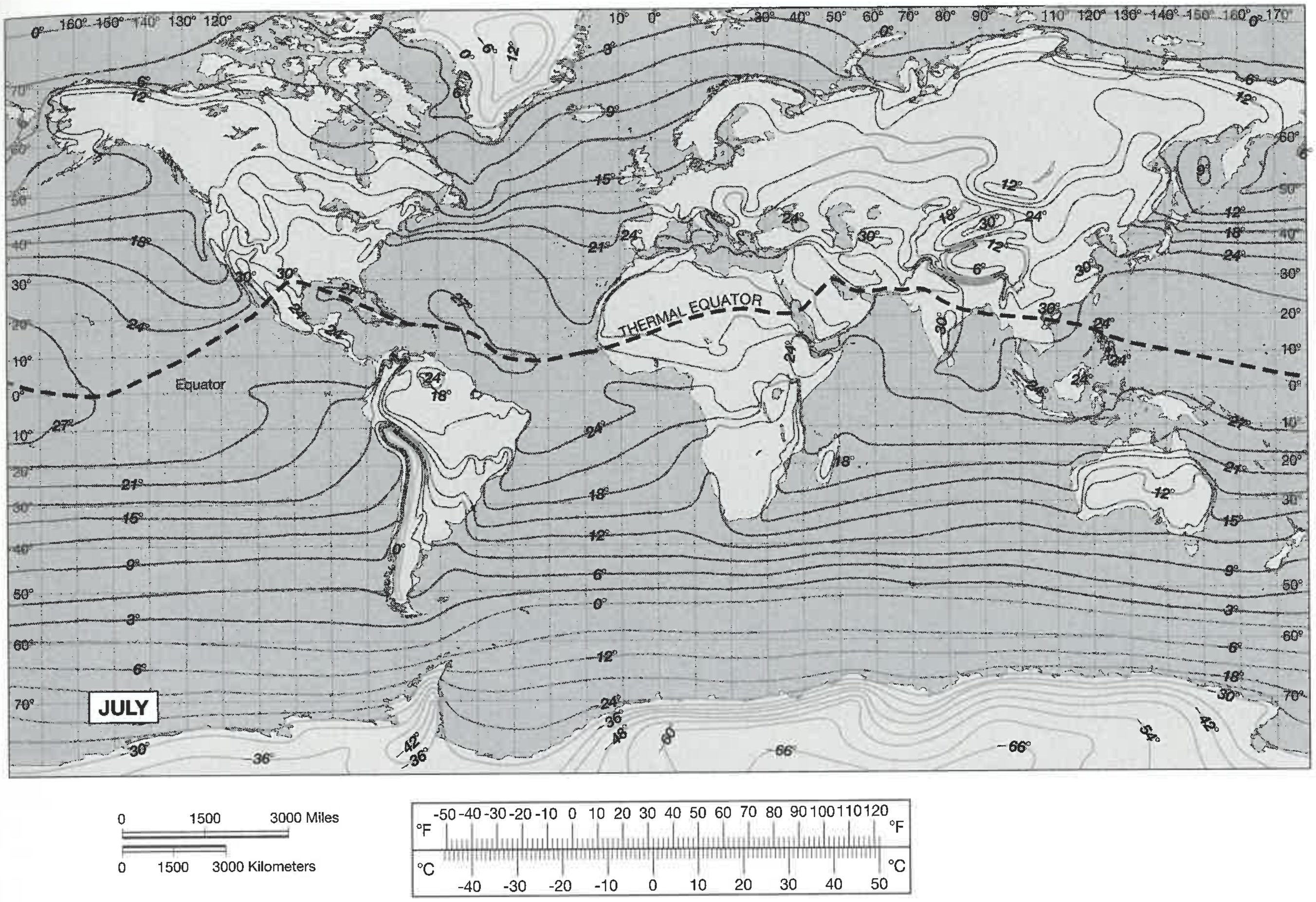
We see, therefore, that the broad pattern of global temperatures is determined by the latitudinal distribution of net radiation, so higher temperatures occur in the tropics and lower temperatures at the poles. This distribution varies with season such that the seasonal range of temperatures is slight in the tropics and increases poleward. Beyond this, we see from the preceding discussion that the seasonal variability is strongly modified by land–ocean contrasts, the interior of continents having a much greater seasonal range than do coastal locations.

As we noted earlier, the circulation systems described here are very generalized; they represent averaged conditions with much of the day-to-day variability

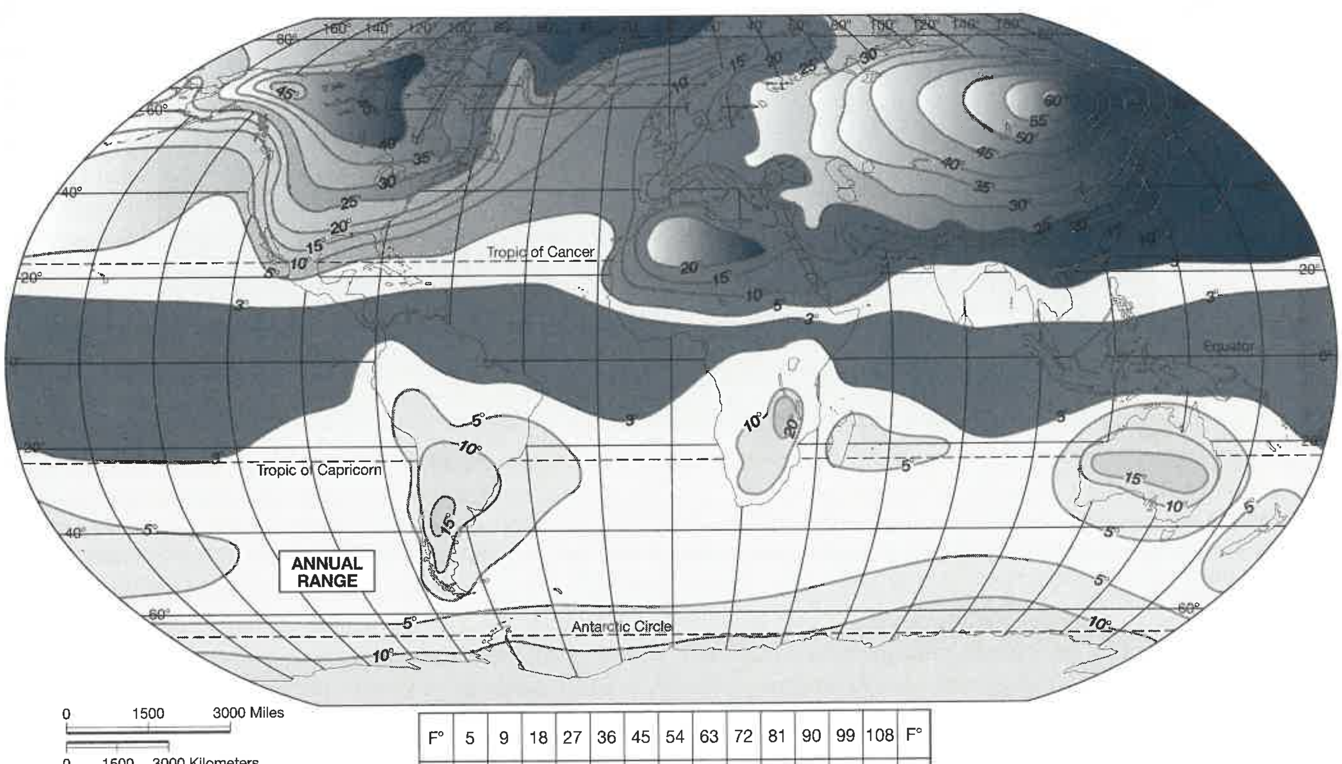


(a)

FIGURE 4-18 Global temperature distributions in degrees Celsius for (a) January, (b) July, and (c) the annual range (difference between summer and winter). (*Source:* From R. W. Christopherson, *Geosystems: An Introduction to Physical Geography*, 3/e, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)



(b)



F°	5	9	18	27	36	45	54	63	72	81	90	99	108	F°
C°	3	5	10	15	20	25	30	35	40	45	50	55	60	C°

(c)

FIGURE 4-18 *continued*

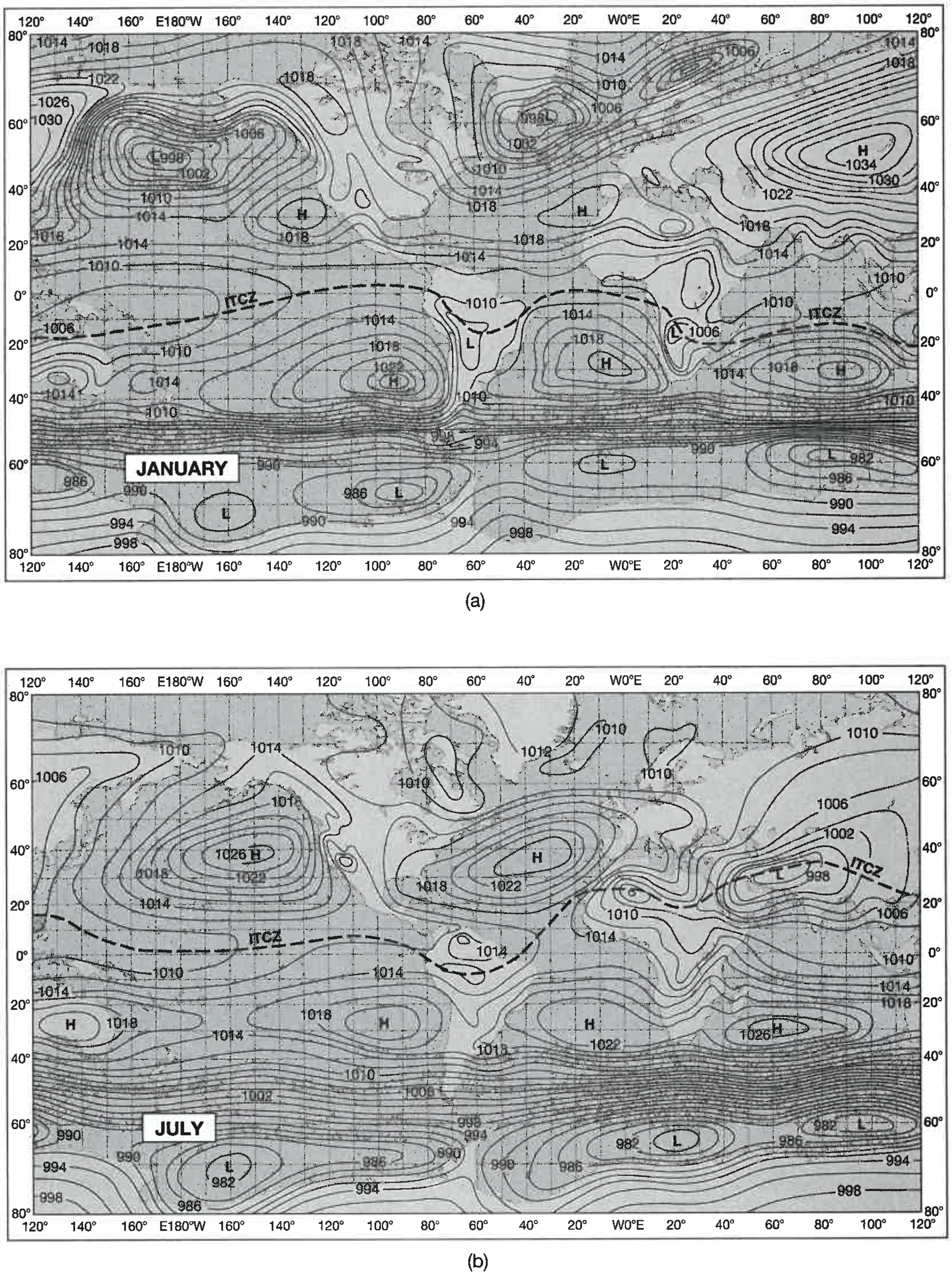


FIGURE 4-19 Average sea-level pressure patterns: (a) January and (b) July. Units are in millibars (mbar), where one atmosphere (atm) equals 1.013 bar (1013 mbar). Low-pressure areas are designated by L and high-pressure areas by H. (*Source:* From R. W. Christopherson, *Geosystems: An Introduction to Physical Geography*, 3/e, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

removed. On any particular day, the circulation and the resulting wind field will be much more complex. As an example, Figure 4-20 shows the surface wind field over the Pacific Ocean measured by a satellite-borne radar system on a September day. The white arrows show the wind direction. The ITCZ is clearly seen where the northeast and southeast trade winds converge just north of the equator. Note that the Southern Hemisphere trade winds are blowing from the southeast—as they move equatorward toward the ITCZ they curve to the left. As they cross the equator

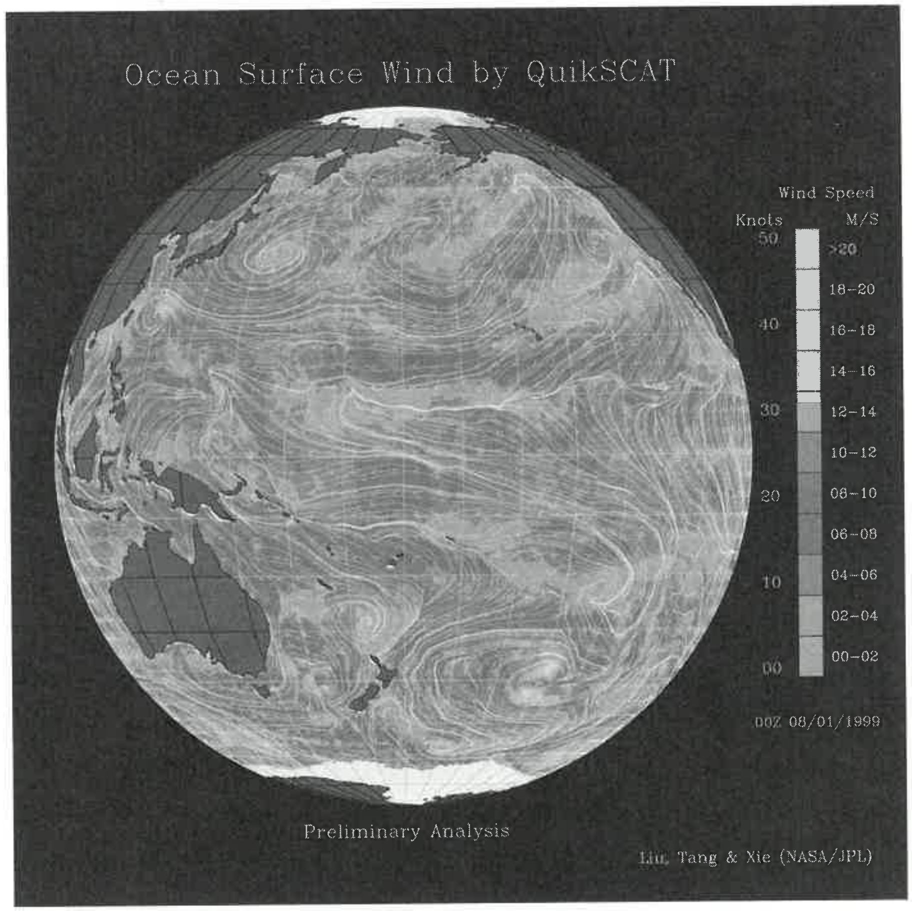


FIGURE 4-20 [See color section] The surface wind field over the Pacific Ocean. The data were derived from a satellite-borne radar system and the white arrows show the direction of air movement at 00Z on August 1, 1999. (*Source:* NASA/Jet Propulsion Laboratory.)

(because the ITCZ is located north of the equator at this point) they move northward and, as they move away from the equator and come under the influence of the Coriolis effect again, they recurve toward the right. This is more apparent in the eastern Pacific. The subtropical high-pressure cells are also well depicted—you can clearly see the air spiraling out from the center of the cells.

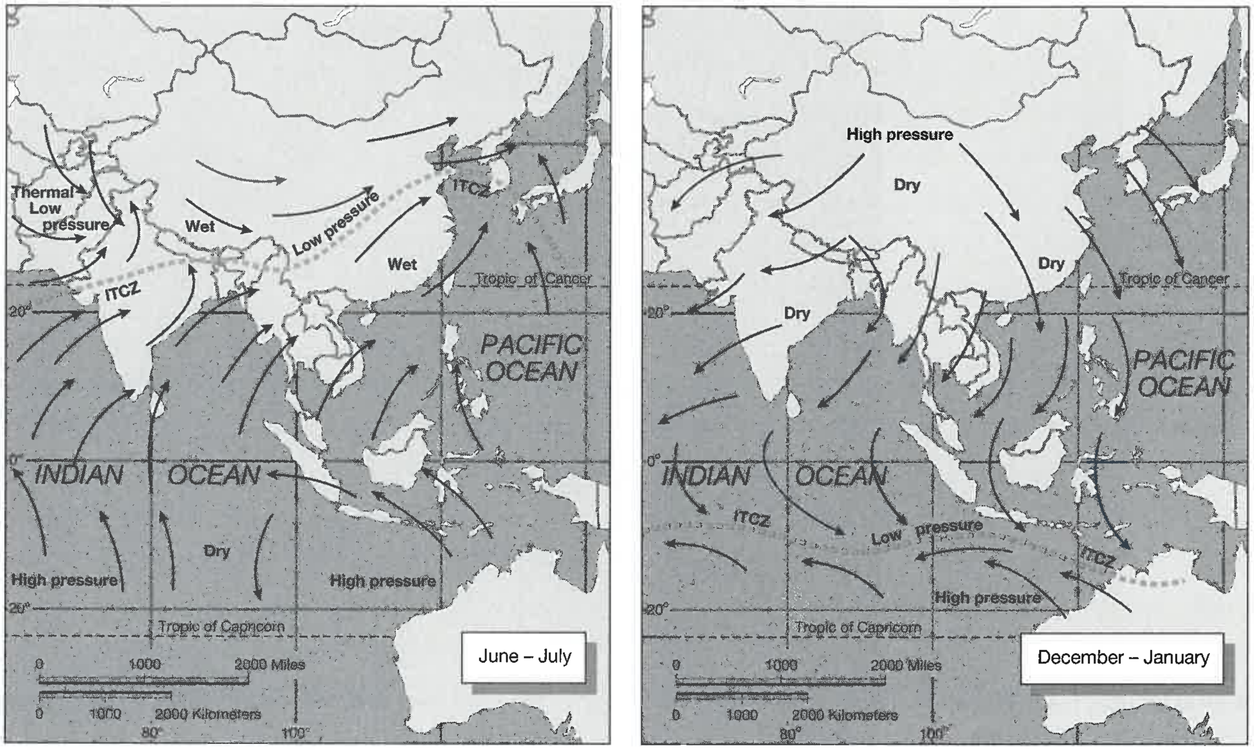
MONSOONS The most extreme consequence of the seasonal variability due to differential heating of land and ocean surfaces is the monsoon regime of Southeast Asia. The **monsoon** is a seasonal reversal in the surface winds. In summer the large Asian landmass, with its high elevations in the Tibetan Plateau of central Asia, causes high surface temperatures, low atmospheric pressures, and intense convection of air above the surface. The rising air is replaced by air moving in from the high-pressure region over the Indian Ocean to the south (Figure 4-21a). The moist air drawn in from the Indian Ocean cools as it rises above the mountains of southwest India and over the Himalayas. In both instances the rising air produces clouds and heavy rainfall (the monsoon rains). In winter the pattern reverses: High elevations and persistent snow cover enhance the continentality, producing even lower temperatures. This results in high atmospheric pressure and subsidence of air over the continent and a southward flow of air (Figure 4-21b). A similar feature, but on a much smaller scale, is found in the southwestern United States, where a "monsoonal" flow from July through mid-September brings moist air in from the Gulf of California and the eastern Pacific.

Global Precipitation Patterns

In addition to transporting energy, the circulation of the troposphere also involves the movement of material across Earth's surface. We can imagine how effective the atmosphere is at transporting material from the fact that pollution from midlatitude industrial sources has been found on both polar ice caps.

The most important substance transported in the atmosphere is water, in the form of water vapor and clouds. Both water vapor and clouds are important for several reasons: They play a dominant role in the global energy balance, they are a significant factor in determining the distribution of freshwater around the globe, and they are highly variable in time and space (making them difficult to predict).

THE GLOBAL HYDROLOGIC CYCLE Water is the most important chemical compound not only in the atmospheric circulation but also in the entire Earth system. The human



(a) (b)

FIGURE 4-21 The monsoon flow over southeast Asia. (a) Summer heating of the Tibetan Plateau produces intense convection and low surface pressures, drawing in moist air from the Indian Ocean to the south. (b) The reverse occurs in winter, when low temperatures and extensive snow cover on the plateau produce high surface pressures, subsidence, and outflowing air. (*Source:* From R. W. Christopherson, *Geosystems: An Introduction to Physical Geography*, 3/e, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

body is 60% water by weight, and all organisms require some water in order to live. From space we see that Earth is a planet dominated by water (Figure 4-22). Seventy percent of Earth's surface is covered by oceans. The poles are encased in extensive sheets of ice that either float on the surface of the ocean (*sea ice*) or form glaciers several kilometers thick over land. Clouds, which are made of condensed water vapor, swirl across the surface, continuously changing in size, shape, and location but always covering about 50% of the globe at any time. The water that exists in gaseous form (water vapor) also varies in amount across the globe—from near zero over the ice caps to about 7% in the tropics.

Water is unique in the Earth system for a variety of reasons, not the least of which is that it is the only naturally occurring substance that can exist in all three phases (solid, liquid, and gas) at the temperatures found on Earth's surface. Because water changes so readily from one phase to another, it cycles easily among all the system components. In so doing, water plays a vital role in many Earth system processes. Changing from a solid to a liquid or from a liquid to a gas requires a large addition of energy. That energy is stored in the water molecule in the form of latent heat. The **latent heat of vaporization,** which we

FIGURE 4-22 [See color section] Earth, viewed from space at about 37,000 km (23,000 mi), is dominated by water. (*Source:* NASA Headquarters.)

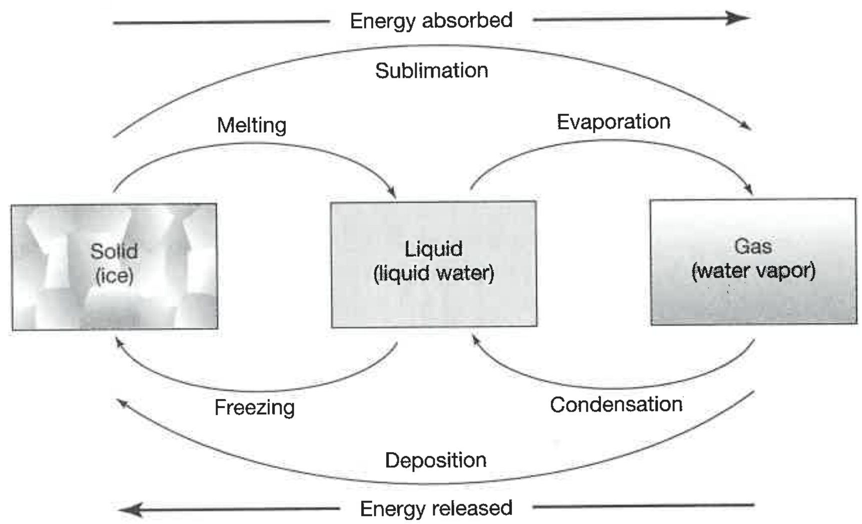


FIGURE 4-23 Schematic diagram of the different phases of water. Energy is absorbed as water changes from a solid to a gas (moving left to right in the diagram) and is released as water changes from a gas to a solid (from right to left).

described in Chapter 3 as the energy needed to convert liquid water to water vapor, is 2260 kJ/kg at 100°C. (This much energy must be added to each kilogram of boiling water to convert it to water vapor.) When the process is reversed and the water changes from a gas back to a liquid, this same amount of energy is released to the environment (Figure 4-23). The **latent heat of fusion,** or the energy needed to convert ice to liquid water, equals 335 kJ/kg at 0°C. When the process is reversed and the water changes from a liquid back to a solid, this amount of energy is again released to the environment. To raise the temperature of liquid water from 0 to 100°C requires 419 kJ/kg. To convert ice to water vapor thus takes 3014 kJ/kg (= 2260 + 419 + 335).

These values apply to water at sea level, and they will vary slightly as atmospheric pressure changes. If these conversion processes occur in different locations, then there is a net transfer of energy from one place to another. Therefore, the distribution and movement of water in its various phases has important consequences for the transfer of energy and the global pattern of surface temperatures.

We saw in Chapter 3 that water vapor and carbon dioxide are the most important of the present-day greenhouse gases. Without them, much of Earth would be too cold to support life. We saw also that clouds have a major impact both on Earth's albedo and on the emission of terrestrial radiation to space. And we will see in Chapter 7, water plays a vital role in breaking down rocks (weathering) and in transporting essential nutrients throughout the Earth system. *Water, in all its phases, is the primary medium by which energy and matter are circulated among the Earth system components.*

Water in the Earth system is concentrated in several major reservoirs:

1. The oceans, where the water exists in the form of seawater;
2. The land surface, in the form of ice sheets, glaciers, snow, lakes, and rivers, and the land subsurface, in the form of groundwater; and
3. The atmosphere, in the form of water vapor and clouds.

These reservoirs and the pattern of water storage and movement throughout the system comprise the global **hydrologic cycle** (Figure 4-24a). Most of Earth's water—about 97%—is stored in the first reservoir, the oceans (Figure 4-24b). Almost 3% is on or in the second reservoir, land. Of this amount, three-quarters is trapped in the polar ice sheets (in Greenland and Antarctica). Should the Greenland ice sheet melt, it would raise the global sea level by about 7 m; should all of Antarctica melt, the sea level would rise by about 57 m. A small amount of water exists in mountain glaciers as well. Most of the remainder occurs as **groundwater,** or water that penetrates through soil and rock and collects below the surface. Water stored in rivers, lakes, and the soil accounts for less than 1% of all the water found on land. Almost two-thirds of this amount is stored in lakes and reservoirs, about one-third occurs in the soil, and a tiny fraction occurs in rivers. The third reservoir, the atmosphere, contains less than 0.001% of all the water on Earth. Figure 4-24b also gives the annual exchange of water among the three major reservoirs.

PRECIPITATION AND SATURATION VAPOR PRESSURE. The transfer of water between the land–ocean surface and the atmosphere takes place through evaporation and *precipitation.* Precipitation occurs when atmospheric water vapor condenses to form small droplets of liquid water. When the water droplets reach sufficient size, they fall because of gravity. If they do not evaporate before they reach Earth's surface, we experience them as rain. If atmospheric temperatures are below freezing, the droplets fall instead as snow or sleet.

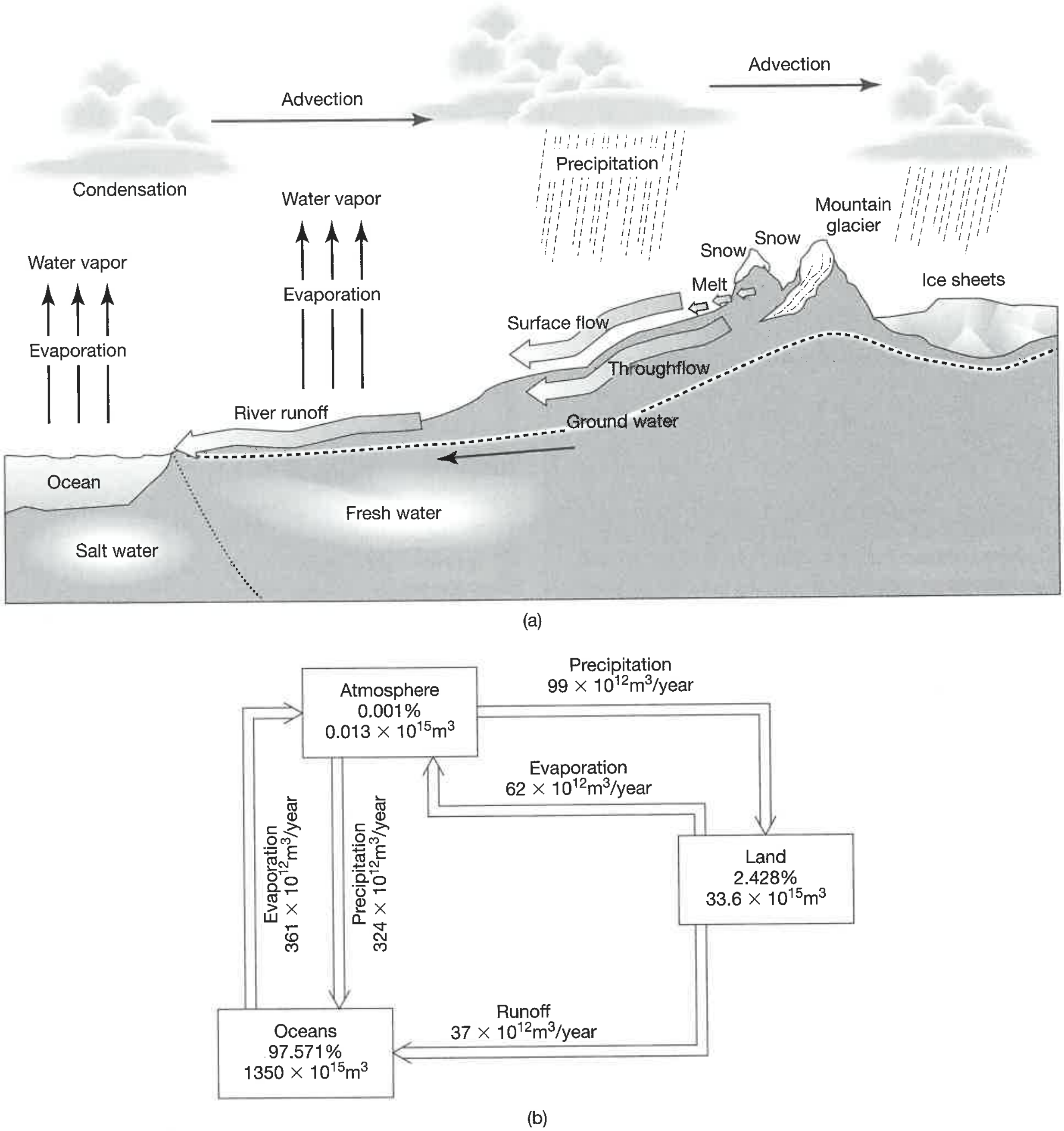


FIGURE 4-24 The global hydrologic cycle. (a) Schematic diagram of how water, in its various phases, is stored and moved throughout the Earth system. (b) The sizes of the major reservoirs of water and the rate at which water is transferred between them. (*Source:* From T. McKnight, *Physical Geography: A Landscape Appreciation*, 6/e, 1999. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

One way of expressing the amount of water vapor present in the atmosphere is to measure the contribution that water vapor makes to the atmospheric pressure. We saw in Chapter 3 that air is composed of numerous different gases. Each gas exerts its own pressure. What is measured as the atmospheric pressure is the sum of all the **partial pressures** of the individual gases—that is, the pressure each gas would exert if it were the only gas present. The pressure exerted by water vapor is referred to as the *vapor pressure*.

Imagine a body of water. Water molecules at the surface that have a little more energy than do their neighbors can overcome the attractive forces that hold the molecules together and thereby escape as water vapor molecules into the air above. This is the process of **evaporation.** Some of these water vapor molecules that subsequently come in contact with the water surface would lose energy, be "caught" by the liquid water molecules, and become liquid water again. This is the process of **condensation.** Once the rate of condensation equals the evaporation rate—that is, as many molecules leave the gas as are added to it—the gas is at equilibrium. At this point, the vapor pressure of water is referred to as the *saturation vapor pressure*. In this scenario, the saturation vapor pressure depends only on the rate at which molecules are transferred from liquid to gas and back again. This rate depends on the energy of the molecules, which means it depends on temperature. (Recall that the higher the energy, the higher the temperature.) Therefore, as temperature increases, the saturation vapor pressure increases.

Figure 4-25 is a graph of saturation vapor pressure versus temperature for water. In general, we can think of clouds as forming when the air is at the saturation vapor pressure for water. Further evaporation adds water vapor molecules to the air, where they condense to form water droplets in clouds. When these droplets become large enough to overcome the upward motion of the air, they fall as precipitation. Assume that an air mass is at the temperature and vapor pressure indicated by point *P* in Figure 4-25. Point *P* is not on the curve; the air is not at the saturation vapor pressure for that temperature, hence clouds would not form and precipitation could not occur. We can bring that air mass to the *saturation point*—that is, to equilibrium at a point on the temperature versus saturation vapor pressure curve—in two ways. First, we could add more water vapor through increased evaporation from the surface and thereby increase the vapor pressure. That action would move the air mass from point *P* up along the vertical dashed line toward the curve. Second, we could reduce the air temperature, which would move the air mass from point *P* to the left along the horizontal dashed line toward the curve.

Because the saturation vapor pressure varies with temperature, knowing just the vapor pressure on a given day does not give a good indication of when clouds will

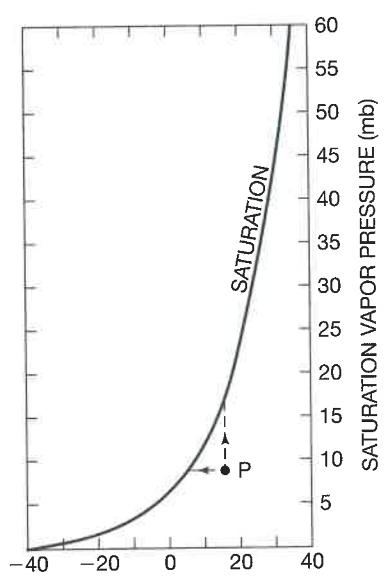


FIGURE 4-25 Saturation vapor pressure versus temperature for water. The curve shows the temperatures and vapor pressures at which the air becomes saturated.

form. Consequently, we normally think in terms of *relative humidity*—the ratio of the actual vapor pressure to the saturation vapor pressure at that temperature. (We are using vapor pressure as a measure of the amount of water present; hence this definition is equivalent to that presented in Chapter 3.) The relative humidity is usually expressed as a percentage; a relative humidity of 100% represents air at the saturation vapor pressure. In general, water vapor will condense to form water droplets and clouds when the air is fully saturated. But in fact, very clean air may have greater than 100% relative humidity (i.e., it can be *supersaturated*) without condensation taking place. Condensation is facilitated by impurities in the air—microscopic particles (solid or liquid) that are small enough to remain in suspension in the air. Such particles are known as *cloud condensation nuclei* (CCNs) when they are used in cloud formation. These nuclei can come from many sources, both natural and anthropogenic. It is likely that many of the clouds that form from CCNs over land surfaces derive from human-produced sources, such as sulfates.

We said earlier that the vapor pressure can be brought to the saturation point either by increasing the vapor pressure or by cooling the air. We can visualize both processes taking place as air moves over different surfaces: evaporation increasing as unsaturated air moves over lakes or the ocean, and temperatures decreasing as the air moves over cooler surfaces. The largest and most rapid changes take place, however, when air rises in the troposphere. As we saw in Chapter 3, temperature generally decreases with altitude in the troposphere, where most clouds exist. Thus most rainfall situations occur with some form of **uplift,** or rising of air masses. Uplift is a general term denoting any process by which air at a given level in the atmosphere is lifted to a higher altitude.

We can get some appreciation for how precipitation is distributed around the globe, therefore, by recognizing that most precipitation takes place as air cools when it is forced to rise. We have already mentioned two of the processes that result in uplift: first, large-scale uplift that occurs with the mixing of air masses of different densities; second, uplift due to convection. Consequently, there is heavy precipitation along the polar front zone in the mid-latitudes and in the vicinity of the ITCZ (Figure 4-26). Convection, however, occurs not only in the tropics, but wherever there is intense surface heating. Therefore, although convection does not always produce rain, it is the dominant rainfall-producing process over warm landmasses in summer. A third process that forces air to rise is the confrontation between a moving air mass and a mountain range. Such encounters cause *orographic* precipitation on the windward (upwind) slopes of mountains. (Orography is the branch of geography that involves mountains and mountain systems.) For example, orographic precipitation commonly occurs on the western slopes of the Sierra Nevada mountain range and on the southern slopes of the Himalayas.

Under these circumstances, precipitation is enhanced due to the atmospheric circulation. Under other circumstances, precipitation is inhibited in certain areas. We call such areas **deserts,** and we can examine why they are located where they are (Figure 4-27). Remembering that condensation results from uplift (which cools the air) or from increasing the amount of available moisture, we would expect to find deserts in areas where uplift is suppressed or where there is an inadequate moisture supply.

In general, precipitation is low in the interior of large landmasses, simply due to the distance from moisture

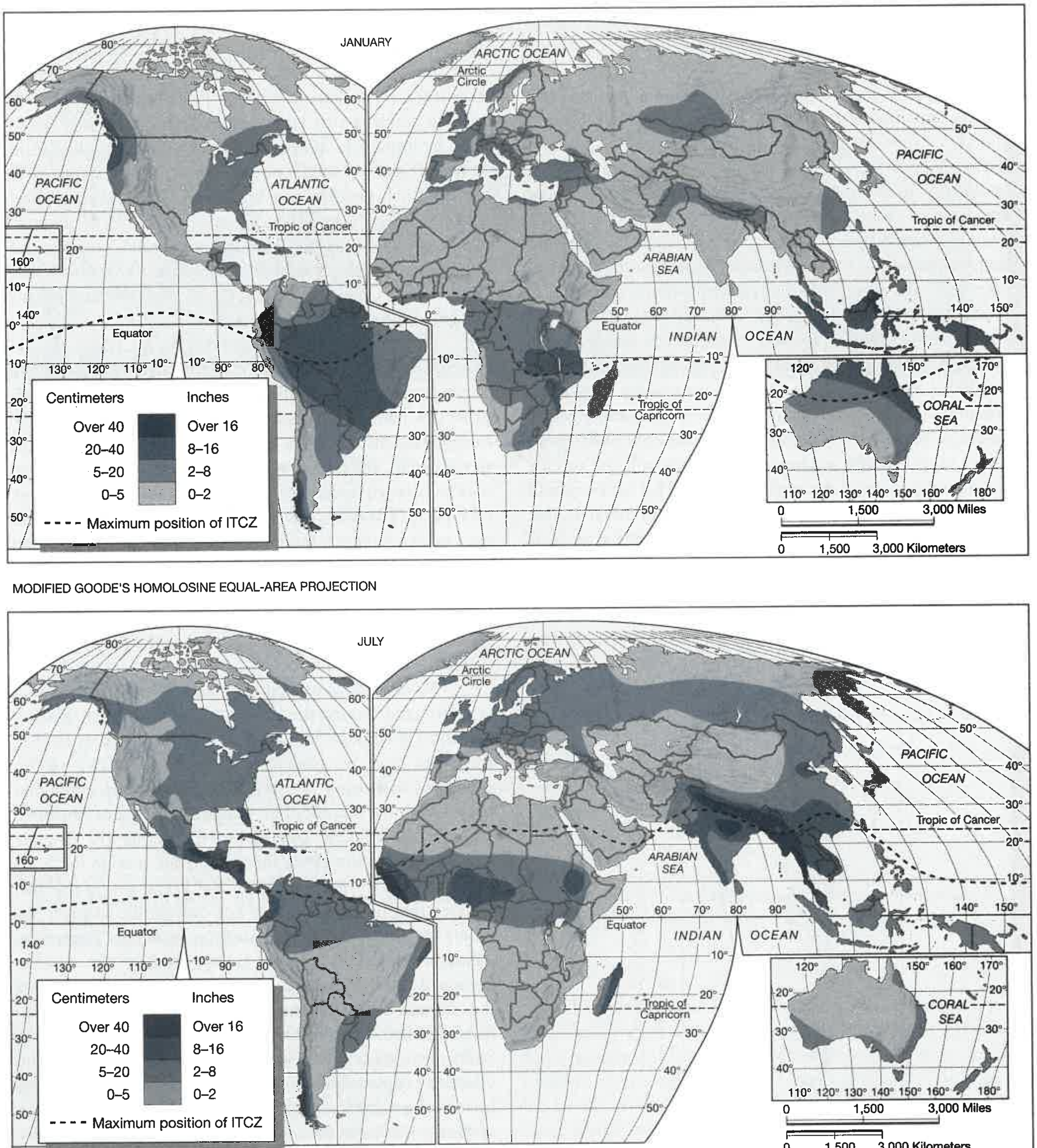


FIGURE 4-26 Global distributions of precipitation over land in (a) January and (b) July. (*Source:* From T. McKnight, *Physical Geography: A Landscape Appreciation*, 6/e, 1999. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

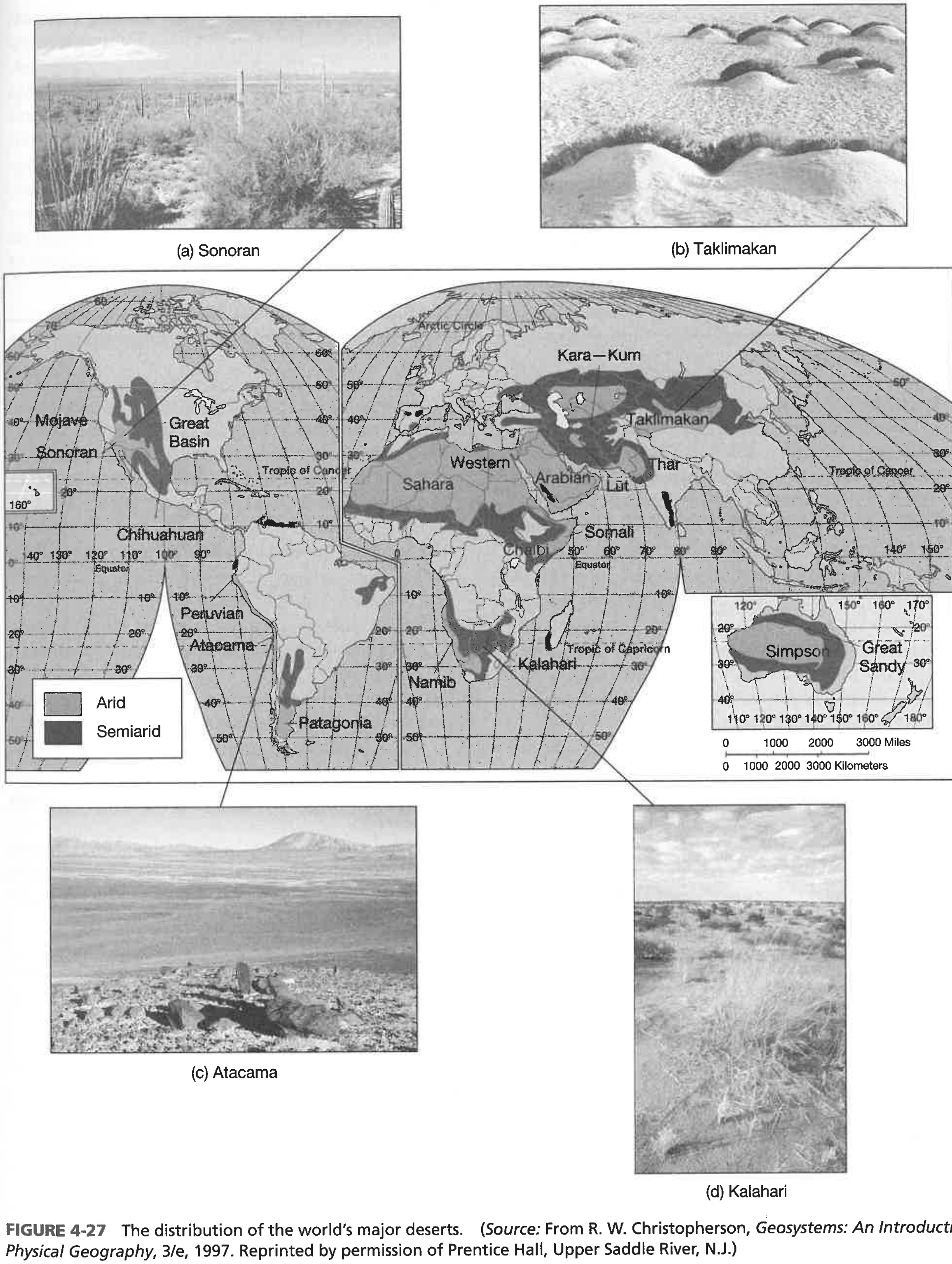


FIGURE 4-27 The distribution of the world's major deserts. (*Source:* From R. W. Christopherson, *Geosystems: An Introduction to Physical Geography*, 3/e, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

supplies. Deserts are located in the vicinity of the descending arms of the Hadley cells, as we noted earlier, and on the leeward (downwind) slopes of mountains. They occur also, perhaps unexpectedly, on the west coasts of large continents in areas that lie equatorward of the midlatitude low-pressure systems. For reasons that we will discuss in Chapter 5, these regions are characterized by cold offshore ocean currents. The cold ocean currents reduce evaporation and cool the air that moves over them, a combination that inhibits convection and precipitation over the adjacent coastline. The deserts that form in this way are called *littoral* (alongshore) deserts. In fact, one of the driest deserts in the world is the Namib Desert along the coast of southwest Africa. The desert of Baja, California (actually in Mexico), is another example. Although you generally think of deserts as being hot, some deserts are also located in the cold polar regions. The low temperatures inhibit uplift, and, where precipitation does occur, quantities are small because of the low saturation vapor pressures. The central part of Antarctica is, in fact, a desert. There is enough ice on Antarctica to raise global sea levels by almost 60 m, but the average annual snowfall on the plateau is the equivalent of less than 51 mm (2 in.) of liquid water.

Chapter Summary

1. The driving force for the atmospheric circulation is the global distribution of energy.
   a. The angle at which the Sun's rays strike Earth changes from the equator toward the poles. The result is that incoming solar radiation decreases with latitude. More solar radiation is received in the tropics than at the poles, resulting in an equator-to-pole temperature gradient.
   b. This temperature gradient drives the atmospheric circulation because of the inverse relationship between the temperature and the density of a gas: Higher temperatures produce lower densities. Differences in the distribution of global temperatures cause differences in air density and, therefore, pressure.
   c. Air tends to move from areas of high pressure to areas of low pressure. These large-scale movements of air produce the global wind belts.
   d. These wind belts are significantly modified by the Coriolis effect, which is caused by Earth's rotation.
2. The net effect of these atmospheric movements is to redistribute available thermal energy. There is a negative feedback between the energy gradient and the circulation. Energy is moved from low latitudes, where Earth is hot, toward the poles, where it is cold.
3. The atmospheric circulation exerts a major control on global temperature patterns. The movement of the air carries water vapor from one region to another. Because evaporation and condensation are largely a function of temperature, the redistribution of water around the globe is also strongly tied to temperature distributions and to the atmospheric circulation.
   a. Precipitation is enhanced wherever the circulation promotes uplift and is inhibited in areas dominated by subsidence.
   b. Precipitation amounts are also affected by continentality and the distance from moisture sources.
   c. The distribution of land and ocean affects the distribution and variability of surface temperatures. Variability increases as the distance from the ocean increases.
   d. The circulation, temperature, and precipitation distributions are modified by seasonal variations in incoming solar energy caused by Earth's obliquity and Earth's orbit around the Sun.

Key Terms

Boyle's law
buoyancy
Charles's law
condensation
convergence
Coriolis effect
deserts
divergence
evaporation
groundwater
Hadley circulation
hydrologic cycle
intertropical convergence zone (ITCZ)
latent heat of fusion
latent heat of vaporization
monsoon
obliquity
partial pressure
polar front zone
subsidence
uplift

Review Questions

1. What are the functions of the global circulatory system?
2. Explain why the distribution of solar energy varies with latitude.
3. a. Draw a graph showing the variation of incoming solar energy and outgoing infrared radiation with latitude.
   b. Indicate the regions of energy surplus and energy deficit.
   c. Explain why this distribution is important for the atmospheric circulation.
4. Explain why heating an air mass causes it to rise.

5. Use a diagram to describe Hadley cells. Why does the Hadley circulation change seasonally?
6. What is the Coriolis effect? How does the Coriolis effect help determine the global pattern of winds?
7. Explain why Earth experiences different seasons throughout the year. Which parts of Earth experience the greatest seasonal variability, and which parts experience the least? Explain why.
8. Contrast the different roles of turbulent heat transfers and conduction in modifying the thermal response of a land surface and an ocean surface.
9. Use map sketches to explain the processes that drive the Southeast Asian monsoon.
10. What is latent heat? Explain why latent heat is important for the redistribution of energy.
11. a. What is meant by saturation vapor pressure?
    b. Draw a graph that plots saturation vapor pressure as a function of vapor pressure and temperature.
    c. Explain why the information shown in the plot is useful for understanding the relationships between atmospheric circulation and precipitation.
12. Describe three processes that produce uplift in the atmosphere and are important in causing precipitation.

Critical-Thinking Problems

1. Sketch a map of India. Locate the major mountain ranges. Show which areas you think would have high rainfall and which areas you think would have low rainfall, and explain why.
2. In this chapter, we discussed several types of deserts, including polar deserts. The center of the Antarctic ice sheet, for example, receives very little precipitation each year and is regarded as a desert, although it does not match the customary idea of what a desert is. From this chapter we saw that:
   - Precipitation generally decreases as temperature decreases (because saturation vapor pressure is much lower in cold air than in warm air).
   - Much midlatitude and high-latitude precipitation occurs in extratropical storm systems that move along the polar front zone.
   - The polar front zone is located in the latitudes where the temperature gradient is greatest. (This will be equatorward of the ice margin, where cold air draining off the ice cap moves equatorward to meet the warm air that is blowing poleward from the subtropical highs.)

   a. Put this information together in a systems diagram that has two feedback loops: one that links ice extent, albedo, temperature, and snowfall; and one that links ice extent, temperature, the location of the polar front zone, and snowfall.
   b. Are these feedback loops positive or negative?
   c. What implications do the feedback loops in part (a) have for the long-term growth of an ice sheet?
3. Indicate on two world maps the areas where you would expect to find relatively high rainfall and where you would expect to find relatively low rainfall, or even deserts, in (a) July and in (b) January. (c) Explain these distributions.

Further Reading

General

Robinson, P. J., and A. Henderson-Sellers. 1999. *Contemporary climatology*. 2nd ed. Harlow: Longman.

Advanced

Hartmann, D. L. 1994. *Global physical climatology*. San Diego: Academic Press.